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GEOLOGICAL MAGAZINE

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Cameral Deposits in Orthoconic Nautiloids

By ROUSSEAU H. FLOWER

ABSTRACT

Previous investigations of the cameral deposits of nautiloids are reviewed, with a discussion of their morphology, their distribution in shells, and growth relationships with other shell parts. Cameral deposits are developed commonly in the orders Actinoceratida, Michelinoceratida, Discosorida, and Oncoceratida. While the patterns and growth relationships of cameral and siphonal deposits are specific, and therefore of great potential taxonomic value, the presence of such structures alone is not a valid basis for the recognition of species or genera.

INTRODUCTION

ALTHOUGH the organic nature of cameral deposits in orthoconic nautiloids was pointed out long ago by Woodward (1851) and reaffirmed by Blake (1882), these structures have never come to be generally known and recognized by palaeontologists. Comments of the earlier palaeontologists have been largely summarized by Holm (1885, pp. 18–19) and need not be reviewed here in detail. Many of these early palaeontologists were impressed by the pseudosepta, really the surfaces of the cameral deposits, and interpreted them as membranes, rather than surfaces of originally solid shell structures. Common also was the misconception that cameral deposits or pseudosepta were formed on the anterior side of the septa when they formed the base of the living chamber, and that they were deposited before the animal moved forward in the shell and secreted a new septum. Curiously, this misconception was expressed by Schröder (1888) and independently by Miller, Dunbar, and Condra (1933). Barrande (1855) recognized the presence of organic deposits in the camerae, but was puzzled as to how they could have been formed. His monumental *Système Silurien du Centre de la Bohême* illustrates many fine sections exhibiting cameral deposits with amazing fidelity, but his comments are largely confined to the phenomena of the deposition of inorganic calcite within the camerae, and he did not distinguish clearly between organic and inorganic structures.

For many years after Barrande's work the study of these structures

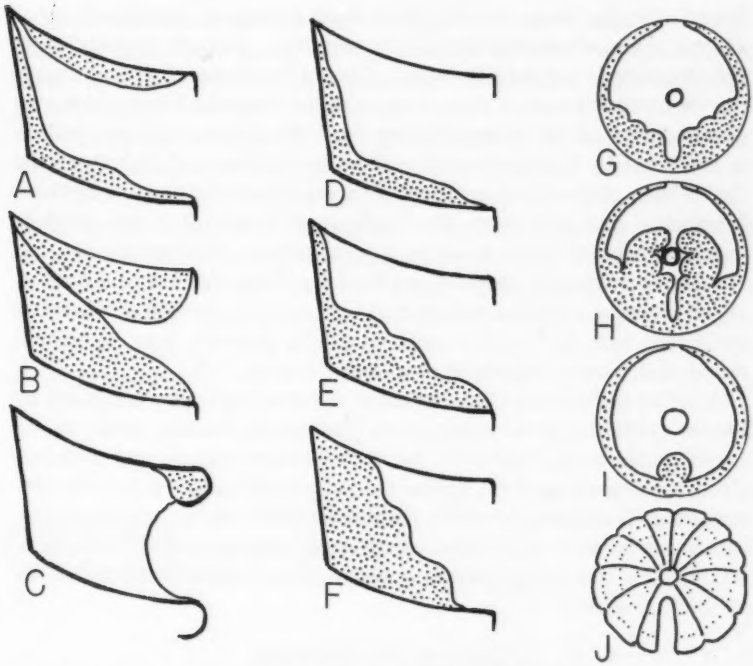
was almost completely neglected, though some palaeontologists, notably Holm (1885) and Ruedemann (1906) called attention to their organic nature. This was reaffirmed by Teichert (1933) in his important study of the actinoceroids. He showed the essential differences between primary deposits of organic origin and secondary deposits of inorganic origin ; differentiated and named the various types of organic cameral deposits (intracameralablagerungen) ; and was the first to emphasize the important role which they played in the hydrostatic adjustment of the organisms.

Flower (1939) dealt with the structures further, proposing some slight modifications of Teichert's terminology, mainly little more than a simplification of an English translation of Teichert's terms. He showed that the cameral deposits are always delayed considerably in their development, being always absent in a series of camerae near the living chamber ; discussed their ontogeny ; and suggested that they were secreted upon the surfaces of cameral tissues, which functioned as mantles, and which maintained metabolic contact with the rest of the animal through the siphuncle. Later (Flower, 1941) impressions of vascular tubes on the surfaces of cameral deposits were described.

Morphology of Cameral Deposits

The essential morphological features of cameral deposits are summarized in Text-fig. 1. Vertical sections through the camerae containing such structures show that deposits may be formed in two main regions : one against the apical septum, both against its free part and its mural part (episeptal deposits, Teichert, 1933), the other against the convex surface of the anterior septum (hyposeptal deposits, Teichert). When both are present and growth is advanced their surfaces may meet along a line termed the pseudoseptum (Text-fig. 1 A–B). A third deposit, possibly a modification of the hyposeptal deposit, is a ring outside the recurved neck of the actinoceroids (Text-fig. 1 C) termed by Teichert the " stutzring ", for which Flower (1939) proposed the name circulus. Episeptal deposits often appear alone without any hyposeptal deposits (Text-fig. 1 D), either concentrated mainly against the free part of the septum (Text-fig. 1 E) or against the mural part (Text-fig. 1 F). When deposits lay mainly against the mural part of the septum they appear to be truncated adorally by the next septum. Their appearance is so different that the term mural deposits was proposed for them (Flower, 1936), though it is recognized that they are only a modification of the episeptal deposits of Teichert.

The distribution of deposits around the periphery of the camerae is not evident in longitudinal sections, though longitudinal vertical sections will often show their marked concentration against the ventral side of the shell. Cross-sections show that they exhibit a bilaterally



TEXT-FIG. 1.—Morphology of cameral deposits. Figs. A–F represent only the left sides of camerae as seen in longitudinal section.

A. Hyposeptal deposits above, episeptal deposits below.

B. Same, in advanced growth stage, deposits meeting along a pseudoseptum.

C. Circulus, surrounding and supporting the recurved septal neck.

D. Episeptal deposits alone, in early stage of growth.

E. Advanced growth stage of true episeptal deposit, with material thickened against free part of septum.

F. Mural deposits, the thickening of the deposit being mainly against the mural part of the septum, and apparently truncated by the anterior septum.

G–I. Cross-sections, showing peripheral distribution of cameral deposits.

G. Early stage of cameral deposit in *Pseudorthoceras*.

H. Advanced growth stage of the same genus.

I. Deposits of " *Orthoceras* " *ludlowense* (Miller and Faber).

J. Sketch of apical surface of internal mould of a camera of *Leurocycloceras*, showing pattern of radial tubes, faint growth lines, lobes, and the impression of a large ventral process of the cameral deposit.

symmetrical pattern, which varies markedly among the genera and species. In general, two main types have thus far been recognized. The Pseudorthoceratidae and many of the Michelinoceratidae (Text-fig. 1 G, H), which lack deposits mid-dorsally, show two thin dorso-

lateral bands, large lobed ventro-lateral masses, separated by a ventral sinus, where the deposit is very thin. In early growth stages the structure is regular (Text-fig. 1 G), but in advanced growth stages the ventrolateral masses may produce rather irregular lobes which may almost surround the siphuncle (Text-fig. 1 H). Quite a different pattern is exhibited by "*Orthoceras*" *ludlowense* (Miller and Faber) of the Ordovician of the Cincinnati region (named from the town of Ludlow, Kentucky, and not from the Ludlow of England); this exhibits instead a deposit which develops a large bulbous midventral mass, but is otherwise thin and simple (Text-fig. 1 I). Dolomite internal moulds of the camerae of *Leurocycloceras* (Flower, 1941) showed a somewhat analogous pattern, but the surfaces of the deposits bear prominent radial and fainter concentric markings (Text-fig. 1 J).

A rather elaborate surface pattern of cameral deposits was figured by Flower (1936) for *Striacoceras typus* (Saemann), showing striae on the dorso-lateral bands, bosses on the ventro-lateral masses, and a ventral sinus, a pattern in general similar to that prevailing in the Pseudorthoceratidae. The cameral deposits may develop elaborate surface patterns, which can be seen only under exceptional conditions of preservation. It is unlikely that our present knowledge of the various patterns is even remotely complete.

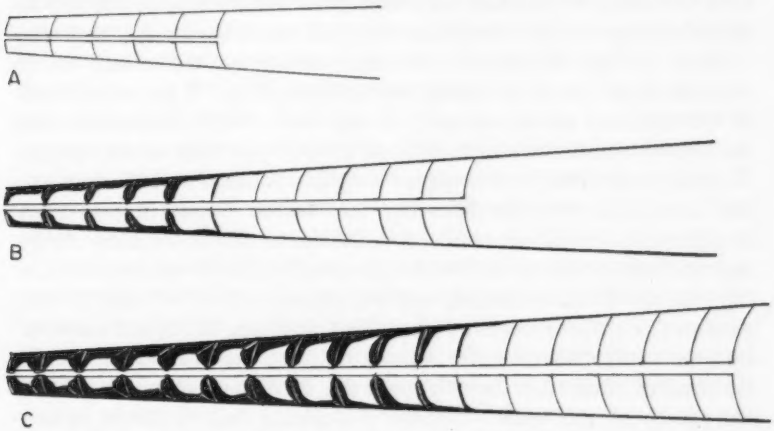
Growth Relationships

The cameral deposits show the same general relationships to the growth of the shell wall and septa as are exhibited by the elaborate deposits found within the siphuncles of actinoceroids and endoceroids. They appear at the apical end of the shell only after a considerable interval, and grow gradually forward. Thus they always lag considerably behind the development of the septa, and there is therefore an anterior portion of the phragmocone which is completely free from cameral deposits. This interval of vacant camerae is constant within the species, varying very little from one growth stage to another. In general, the anterior six to fourteen camerae commonly lack deposits, though the number may be greater in some species and smaller in others.

Text-fig. 2 shows something of the growth relationships; for simplicity the small apical camerae are not represented. Natural truncation of the shell is not implied by this figure; truncation is an exceptional rather than a typical phenomenon. Text-fig. 2 A represents the adoral portion of a young shell. The five illustrated camerae are vacant. In a more advanced growth stage (Text-fig. 2 B) a different anterior five camerae lack deposits, but the apical six camerae possess them. The older apical deposits are more advanced than those at the anterior end. A still more advanced stage (Text-fig. 2 C) shows in the figure only the extreme base of the living chamber. The anterior five

camerae are vacant, as before. The deposits can be traced from the youngest in the sixth camera to the oldest in the seventeenth, and progressively older deposits show greater thickening and more advanced growth stages.

It can be seen at once that small fragments of a mature shell as in Text-fig. 2 C will show no deposits if they represent the anterior part of the phragmocone, immature deposits if they represent the middle portion, and advanced deposits if they represent apical camerae. From the very different appearance of such fragments, one might easily reach the erroneous conclusion that they either represented different species or that cameral deposits are erratic in development within a species.



TEXT-FIG. 2.—Three successive growth stages of a species of *Geisonoceras*, showing growth relationships of cameral deposits to the shell as a whole. The shells grow adorally toward the right of the figure.

The matter may be even more perplexing if one is confronted by commensurate fragments from individuals representing different growth stages of the same species. One can see from Text-fig. 2 how different in aspect are the five camerae of Fig. A from the apical five of Figs. B and C. It should be noted in this connection that abundant material has substantiated the growth relationships postulated above, and has not confirmed the view of Miller, Dunbar, and Condra (1933) that cameral deposits were formed at the base of the living chamber.

ECOLOGICAL SIGNIFICANCE

The cameral deposits of the Michelinoceratida serve a very important function in maintaining the balance of the animal in life. They weigh down the apex of the shell, permitting the animal to live in a horizontal rather than a vertical position. Their concentration on the ventral side

stabilizes the shell with the venter beneath, and the heavy tip of the shell must have been a material aid to the animal when it swam by means of the hyponome. The deposits within the siphuncles of the Michelinoceratida are small. Even in the actinoceroids, where the siphuncle is large and its apical segments may be mostly filled by annulosiphonate deposits, the weight of these deposits played a relatively minor role in the hydrostatic adjustment in contrast to the cameral deposits, as shown by Teichert (1933), who estimated the mass and weight of these various parts.

In the endoceroids the large siphuncles filled with endocones play a similar role. It should be emphasized that the siphonal deposits of both the actinoceroids and the endoceroids are delayed in relation to the development of the phragmocone, much as are the cameral deposits.

Prior to the recognition of these structures there was much uncertainty as to how the straight cephalopods lived. If gas was present in the camerae, the apical part of the shell would be buoyant, the anterior end would be weighted down by the body mass of the animal. The shell would then have occupied a vertical position with the apex up. This conflicted with the discovery that colour bands in numerous straight cephalopods are confined to the dorsal side of the shell, showing conclusively that the animal held the shell in a horizontal position in life with the dorsum upward, and led Schmidt (1930) to suggest that perhaps the anterior camerae were filled with gas, the apical camerae with sea-water. Many other authors doubted the presence of gas in the camerae. Indeed, it was this doubt that is responsible for the general acceptance of the term "camera", implying only a closed space, instead of the term "air chamber", employed in the earlier descriptions of cephalopods.

Those who postulated that gas was absent in the camerae were immediately faced with a new dilemma. If there was no gas in the camerae, what could have been the function of the long phragmocones developed in many straight cephalopods? There were hundreds of species of straight cephalopods, from their origin in the Ordovician to their extinction in the Triassic, many of which attained immense size, most of the length being devoted to the phragmocone. The endoceroids of the Ordovician reached a length of considerably more than 12 feet, with 11 feet of that length devoted to the phragmocone. The actinoceroids of the Ordovician attained lengths of 2 to 4 feet, and the Mississippian *Rayonnoceras* probably of 6 feet. The Michelinoceratidae were smaller in general, but *Michelinoceras niagarense* is known to have attained a length of 5 feet, and there are Middle Silurian species in Bohemia which were as large. It is impossible to believe that such shells could have developed if the phragmocone did not offer some very definite advantages.

It is the deposits of the camerae and siphuncle which supply the solution to this problem. They add weight to the apex of the shell ; their mass reduces the space available for gas in the apical camerae. Therefore the centre of gravity is shifted backward, the centre of buoyancy is shifted forward in the shell. It is only conjectural, but certainly possible, that some species were so adjusted that the weight of the animal was very nearly equal to the weight of the volume of water it displaced. Such forms could have been active swimmers. Others, with heavier and more strongly developed deposits, lived as vagile benthos.

Once this balance was achieved it had to be maintained. Without further growth of the cameral deposits, the balance would be destroyed as the animal added more material to the aperture, moved forward, and secreted more septa, forming more air chambers. To maintain the balance of the shell with further growth, the cameral deposits extended as indicated in Text-fig. 2, thickening in the apical camerae by the addition of more material to their surfaces, and appearing in camerae progressively farther orad, but still maintaining a fairly uniform distance between the last and youngest deposit and the base of the living chamber.

The deposits exhibit surfaces which are often so complex that a precise determination of their mass and weight would be almost impossible to obtain much more closely than has already been done by Teichert (1933), but it is a reasonable inference that the centres of gravity and buoyancy were so maintained at the mid-length of the animal as to permit it to live in a horizontal position.

Origin of Cameral Deposits

Cameral deposits exhibit a definite growth relationship to that of the shell as a whole, and develop bilaterally symmetrical patterns which are characteristic of the species. They show growth lamellae and, under favourable conditions, prismatic material normal to the growth surfaces, a feature characteristic of shell parts, and one due to the presence of prisms of secreted aragonite. These are features of a shell structure and, further, of one laid down by a specialized secreting surface, for which the term cameral mantle was proposed (Flower, 1939). When the animal moved forward in the shell, it may be supposed to have left behind a layer of tissue attached to the anterior wall of the septum and directly connected with the siphonal strand. Later the next septum was formed and the connecting ring was secreted within the wall of the siphonal strand.

Happily, further evidence of such tissue was found. Flower (1943) found structures in *Rayonnoceras* which could only be interpreted as a calcite filling of cameral tissues which were retained in the shell when

sediment invaded the siphuncle and penetrated partially into the camerae through the broken siphuncle wall. In the actinoceroids the cameral tissues were extremely active and must have retained vascular connection with the visceral mass throughout life, to obtain the material for the thick and heavy cameral deposits. The only possible connection is through the connecting ring, which is spicular and therefore minutely porous. No pores large enough to admit possible macroscopic blood tubes could be found. There must therefore either be a penetration of the ring by capillaries or the exchange of blood from cameral to siphonal tissue through a haemocoel lying just within the connecting ring. To facilitate the transfer of blood between the camerae and siphuncle the actinoceroids developed greatly expanded siphuncle segments (increasing the surface area of the connecting ring) and a system of elaborate tubes, a central canal, a series of radial canals, terminating in a hollow space, the perispatium, just within the connecting ring (Teichert, 1933). These structures can be interpreted convincingly only as blood tubes. Further, in the evolution of the actinoceroids there is a specialization of these tubes : in *Polydesmia* the radial canals are numerous and irregular ; in the Armenoceratidae they are developed into a series of double arcs ; finally into a series of single arcs in the Actinoceratidae and most Ormoceratidae, and the radial canals may even show in themselves a bilaterally symmetrical pattern (Flower, 1940). In every case the system terminates, as far as we can tell, in the perispatium within the connecting ring. Either this may represent a region at which a haemocoel is developed or a region in which the blood-vessels divide into capillaries and, as such, continue into the cameral tissues ; other considerations (Flower, 1939) indicate that the hypothesis of continuous capillaries is slightly preferable.

If such large and regular blood vessels can develop in the siphuncle of the actinoceroids, is it possible that in some nautiloids where the connecting ring was absent or poorly calcified, and thus did not form a barrier to large tubes, a similar specialization could have occurred within the cameral tissues ? Such thick-walled tubes in a thin tissue would form impressions on the surfaces of the cameral deposits lying against them, and would then be represented by tubular grooves on its surfaces. Impressions of just such tubes were found in the dolomite fillings of the camerae of *Leurocycloceras* (Flower, 1941), and their occasionally serpentine course, as well as the bilateral symmetry of the pattern which they formed, left no doubt as to their organic nature, and very little as to their interpretation.

Taxonomic Use of Cameral Deposits

The varied patterns shown by cameral deposits are characteristic of species and of genera. Unfortunately, so little is known of their

details in such a large number of species and even genera that they are not at the present time highly useful in taxonomy. There are instances where such deposits have been misunderstood and generic names have been based upon them.

The genus *Arthrophyllum* was proposed by Beyrich (1850) for a remarkable object which he believed to be a coral. It was a portion of a straight cephalopod showing peculiar radial lamellae which are clearly cameral deposits. Roemer (1852) later figured this material, identifying it with the species which he had formerly (1843, p. 35, pl. 10, figs. 6–10) described as *Orthoceras crassum*. This species is therefore the type species of *Arthrophyllum*. Dahmer (1939, pp. 340–7) restricted the species *crassum* to the original of Roemer's (1843) pl. 10, fig. 6, a portion of a straight cephalopod showing no evidence of the cameral deposits. One might conclude that *Arthrophyllum* does not demonstrably possess the deposits on which it was originally based and may, from Dahmer's restriction, be considered a synonym of *Michelinoceras*. However, the regional distribution of the deposits, their clear development only in camerae considerably removed from the living chamber, were not understood by Dahmer. It may well be that he was incorrect in dividing *Orthoceras crassum*, and that the apical portion of a phragmocone which shows cameral deposits, to which he gave the name *Orthoceras kahlenbergensis*, could be a fragment representing the early portion of a relatively mature specimen of *Orthoceras crassum*. This can be determined only by a study of more complete material, preferably from the type locality. If this supposition is true, *Arthrophyllum* Beyrich may prove to be sufficiently distinct to merit recognition. Certainly, what little is known of its cameral deposits indicates a pattern somewhat different from the patterns known from other orthoconic cephalopods.

Grabau (1922) based the genus *Stereoplasmoceras* primarily upon the presence of heavy cameral deposits in straight cephalopods. He had included in his genus some actinoceroids which have since been removed to their proper position, but the generic name has been restricted (Kobayashi, 1936) and the genus *Stereoplasmoceras* is valid.

LeMaitre (1950) called attention to orthoconic cephalopods from North Africa showing structures reminiscent of those of corals, and compared her material, described as *Orthoceras coralliforme*, with *Orthoceras kahlenbergense* Dahmer. Her illustrations show a series of short portions of phragmocones from which the cameral deposits have been largely removed along with the shell wall and mural part of the septum. They are strongly reminiscent of the cameral deposits figured for *Striacoceras typus* (Saemann) (Flower, 1936), but show a more finely sculptured pattern. It is not evident that all of the specimens figured are conspecific ; in fact it is eminently possible that when these

forms are better known *Orthoceras coralliforme* may prove to include representatives not only of several species, but of species falling into several genera and families.

H. and G. Termier (1950) have, however, erected a new genus *Lamellorthoceras*, and two new species, *L. gracile* and *L. vermiculare*, illustrated by rather crude drawings, but evidently representing similar internal moulds of camerae in which cameral deposits are well developed. Adequate criteria for comparison with previously described genera, the structure of the siphuncle, the cross-section, suture, and surface patterns, are unknown. Assuredly, while it is important to note the presence of specimens from North Africa exhibiting cameral deposits, the features shown by these specimens are not in themselves a good basis for the recognition of new species or of a new genus.

Turner (1951), in his study of *Rayonnoceras*, also seems to have been unaware of the nature of cameral and siphonal deposits, and to be unfamiliar with Teichert's (1933) careful study of these structures in actinoceroids. He apparently regards the cameral and siphonal deposits as having been formed contemporaneously with the formation of camerae, for he notes that they are absent in some adoral camerae close to the living chamber but states : " It appears that in *Rayonnoceras* a calcified siphuncle and septal linings may have ceased to be formed while camerae were still being added to the conch." Here again, a large part of the difficulty stems from the fragmentary nature of many specimens. A portion of a shell not retaining the living chamber will lack deposits if it was located close to the living chamber, but apically will show deposits to an extent depending directly upon distance from the living chamber at the time of death of the animal, as indicated in Text-fig. 2. As a consequence the degree of calcification of the deposits of the camerae and siphuncle are not a valid specific character unless the relation of the portions upon which these observations are based to the living chamber are known. Turner has employed such features as criteria of some of his species. Plainly in some cases the species lacking calcification are based upon fragments consisting only of a few adoral camerae attached to the base of living chambers. In other cases the fragments are only pieces of phragmocones, and the exact relative position of such portions to the whole shell cannot, of course, be demonstrated.

Taxonomic Distribution of Cameral Deposits

As previously noted, cameral deposits are best developed in straight cephalopods, those which a generation ago were placed in the too-comprehensive genus " *Orthoceras* ". It remains to consider their distribution within the taxonomic divisions now in use (Flower and Kummel, 1950).

Cameral deposits are unknown in most of the Ellesmeroceratida. They are developed in members of two of the more specialized families of that order, the Baltoceratidae and the Protocycloceratidae.

Careful restudy of the type and other material has led to the conclusion that the structures of *Vaginoceras oppletum* which Ruedemann (1906) regarded as organic cameral deposits represent only inorganic calcite fillings. Extensive search has failed to reveal organic deposits in the camerae of any of the Endoceratida. Quite probably the hydrostatic role played by the cameral deposits in other cephalopods is taken over in the endoceroids by the endocones filling the large ventral siphuncles. If one employs Teichert's estimates, it becomes probable that the slender endoceroids, even the giant forms of the Middle and Upper Ordovician, may well have been active swimmers. The piloceroids, however, as suggested by both Teichert (1933) and Kobayashi (1936) must have been bottom living.

Cameral deposits are very common and may have been universally present in the Actinoceratida. In some of the smaller members of the Oncoceratidae their development may be so retarded that they are found only in the apical portions of large shells, and may not be apparent from specimens representing considerable portions of the anterior parts of phragmocones.

A more perplexing question is whether they are present in all Michelinoceratidae. They are well developed in many of the large Silurian and Devonian species. The patterns of the deposits of many such shells have been figured faithfully and extensively by Barrande (1865–1870), and are particularly well shown in the large orthocones of Étages e1 and e2 of Bohemia. There are, however, some small, slender shells for which no cameral deposits are definitely known. The study of such species of the American Silurian and Devonian has shown that cameral deposits are developed, but they are often so retarded in their development that they are present only in extreme apical portions of the phragmocones. They are not developed in the later growth stages which, being larger and more conspicuous, constitute the bulk of the material in most collections. I have for some time believed the cameral deposits to be wanting in some Chazyan species, but further study, based upon more complete material, has demonstrated the presence of these structures in many species, and leads me to suspect that the same may prove to be true of others when adequate material can be studied. Typical orthoconic specimens from the upper red and upper grey *Orthoceras* limestones seem to lack deposits, but again I have been able to demonstrate their presence in septate fragments evidently representing apical portions of many of these species, broken off and often deposited together by themselves. No deposits are known in small, extremely slender shells, as *Bactrites*, *Arkonoceras*, *Plagiostomoceras*, or

of the very slender *Orthoceratites gracilis* of the black shales of the Devonian of Germany. If they are absent in such forms, one is again faced with the old dilemma of their mode of life. However, up to the present none of these forms have been examined by sections or otherwise from material complete enough to warrant the conclusion that cameral deposits are completely wanting, except for the anomalous *Bactrites.*

Cameral deposits are known for three of the Ordovician genera of the Discosorida, *Ruedemannoceras*, *Westonoceras*, and *Teichertoceras.* They are doubtfully present in the extreme early stages of *Tuyloceras*, but are unknown in other Silurian genera. It may well be that cameral deposits became superfluous when the siphuncles became large and the growth of their deposits was advanced, as in the Silurian Discosoridae.

In the Oncoceratida cameral deposits are often extremely thin and appear to be present in all camerae of mature specimens. In *Westonoceras* of the Discosorida, internal moulds from which cameral deposits are exfoliated along with the mural parts of the septa and the shell wall, the deposits are often apparent by rather shallow, widely-spaced longitudinal markings on the outer walls of the camerae. Similar longitudinal markings are not uncommon in the Oncoceratida, though they are not by any means found in all species or genera. They are most commonly seen in some of the more specialized Silurian genera in which the aperture of the shell is contracted at maturity, notably the members of the Hemiphragmoceratidae and Trimeroceratidae. They are again often well developed in Devonian breviconic genera, and are strongly developed in the gyroconic Devonian genus *Stereotoceras* which belongs to the Brevicoceratidae (Flower, 1950). In the oncoceroids the cameral deposits lack the hydrostatic function which they plainly have in the dominantly orthoconic orders. Instead, the appearance of cameral deposits is one of several manifestations of maturity. Once the shells develop contracted apertures, further deposition of calcium carbonate as an extension of the shell wall at the aperture or as additional septa is not possible. Excess calcareous material is then secreted in different ways : (1) by a thickening of the interior of the shell near the aperture, (2) by the deposition of a thick band near the base of the living chamber, and (3) by the development of very thin linear cameral deposits, which appear to develop simultaneously throughout the length of the phragmocone, and fail to show any marked thickening in the apical camerae.

True cameral deposits are unknown in the Tarphyceratida. Ulrich, Foerste, Miller, and Furnish (1942), have reported cameral deposits in the Trocholitidae. The structures found there are not analogous to those discussed in the preceding pages. They are not peripheral, and are probably not a primary secretion by a cameral mantle. Rather, they

appear to be outgrowths of the rather thick connecting rings, which may extend out from the siphuncle for some distance along the free parts of the septa. No such structures have been reported in the Barrandeoceratida, Rutoceratida or, in short, any of the younger groups of nautilicones. In these shells the buoyant camerae were located above the living chamber, and there was no hydrostatic need for the cameral deposits as there was in the straight cephalopods.

Natural truncation of the shell is rare in the Nautiloidea. It occurs in the anomalous *Orthoceras truncatum* Barrande, and in the order Ascoceratida. In both cases the shedding of camerae reduces the buoyant effect of the apical phragmocone, the need for cameral deposits as hydrostatic adjustors is lessened or completely removed, and these structures have not been recognized.

Antiquity of Cameral Tissues

Cameral deposits are developed widely in the orders Actinoceratida and Michelinoceratida, and less widely in the Discosorida, specialized families of the Ellesmeroceratida, and in many of the Oncoceratida. Uniformity of the structures in these diverse groups makes it extremely unlikely that cameral tissues developed independently within these orders; rather it is highly probable that they were present in the common ancestors of these forms. Their presence is, however, undetected, because cameral deposits were secreted only in specialized groups where such activity became a physiological or ecological necessity. As outlined elsewhere (Flower, 1954), the latest point at which these diverse groups can be united in a common ancestor is in the family Ellesmeroceratidae, of the order Ellesmeroceratida. This family is almost completely confined to the lower Canadian (Arenig). There this family is joined by some members of the Plectronoceratidae, the other genera of which comprise the only adequately known two genera of the Upper Cambrian which can be regarded as true cephalopods. The cameral tissues may, then, have been archaic features of the cephalopods.

The alternate hypothesis, that cameral deposits developed independently in several different orders is not impossible, but is less convincing. It is difficult to see how, with several independent developments of such tissues, the deposits which they secreted agreed so closely in their general patterns in the different orders in which they appear. For some years palaeontologists were similarly puzzled by the erratic distribution of actinosiphonate structure in the Oncoceratida and Rutoceratida. It finally became evident that actinosiphonate deposits represented excess calcification of the connecting ring, which conformed to a tissue pattern of the siphonal strand which must have been common to most members of the Oncoceratida; the structures were retained in

only a few of the older genera of the Rutoceratida, which stem not from the Barrandeoceratida, the nautiliconic stock dominating the Ordovician and Silurian, but from the Silurian Oncoceratida. It now seems that the cameral deposits are a similar manifestation of cameral tissues which were widely present in the older cephalopods, but which only secreted calcareous materials within the camerae in specialized groups and under specialized conditions.

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Alkali Metasomatism and the Petrology of Some Keratophyres

By M. H. Battey

(PLATES V AND VI)

Abstract

Petrographic and chemical study of New Zealand keratophyres, occurring in a typical spilite-keratophyre association, shows that they were originally partly glassy flow-rocks and breccias comparable with rhyolites of young volcanic areas. They are now composed of a low temperature mineral assemblage inconsistent with textural evidence of high temperature extrusive origin, and alkali metasomatism has enriched some rocks in soda and others in potash, so that their compositions are symmetrically disposed about that of a primary rhyolite magma with roughly equal amounts of the two alkalis.

Data from other occurrences suggest that this condition is common amongst described keratophyres.

Though the rocks do not originate from any special magma, the term keratophyre should be retained, with appropriate qualifiers, for old rhyolites with low-temperature mineralogy, which, incidentally, will often be found to have been affected by alkali metasomatism.

Introduction

THE keratophyres which form the subject of this paper come from three localities in the extreme north of the North Island of New Zealand, Doubtless Bay, Mount Camel, and Three Kings Islands, where they occur interbedded with spilites. Bartrum (1929) described a keratophyre from Mount Camel under the name albitic granophyre and gave a chemical analysis. In 1936 he described three rock types, quartz-keratophyre, albitic porphyry, and spilite from Three Kings Islands, this being the first recognition of spilitic rocks as such in New Zealand (Bartrum, 1936). The field relations of the lavas have been described by the present writer (Battey, 1950, 1951).

Petrography

The keratophyres are of simple constitution, being composed of quartz, with albite and potash feldspar associated with it in widely varying and inversely related proportions, and much smaller amounts of finely-divided ferriferous chlorite or greenish biotite. Dusty leucoxene is almost always present and zircon is a very constant accessory. They contain sparse phenocrysts of the alkali feldspars, but phenocrystic quartz is not found, although in a few cases there are polycrystalline secondary quartz patches. Following the usage of Beskow (1929, p. 82) and Geijer (1931, p. 189) [1] the term quartz-keratophyre is not employed, but is kept for types with phenocrystic quartz.

[1] Geijer used "quartz-bearing porphyry" for types with more than 5 per cent (usually about 20 per cent) of groundmass quartz.

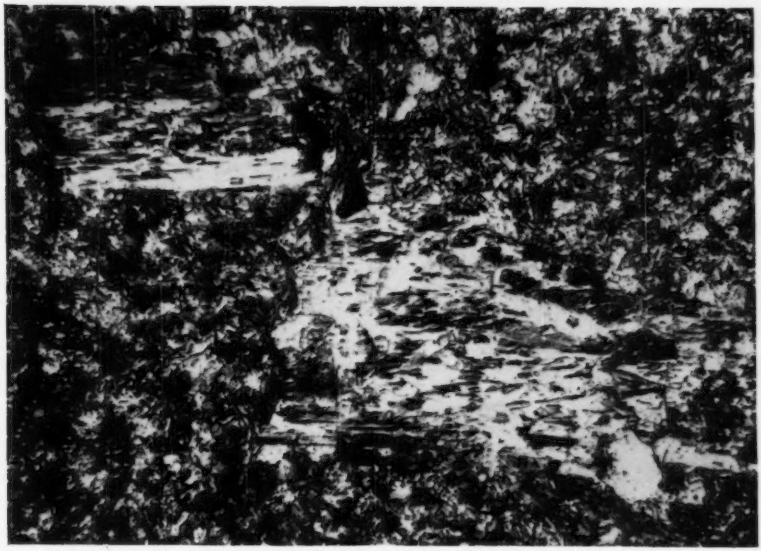


FIG. 1.

VAGUELY-BOUNDED "COMPOSITE" ALBITE PHENOCRYSTS MADE UP OF SLIGHTLY DISORIENTATED TWINNED LATHS, IN KERATOPHYRE, RANGIAWHIA PENINSULA. CROSSED NICOLS, × 33.

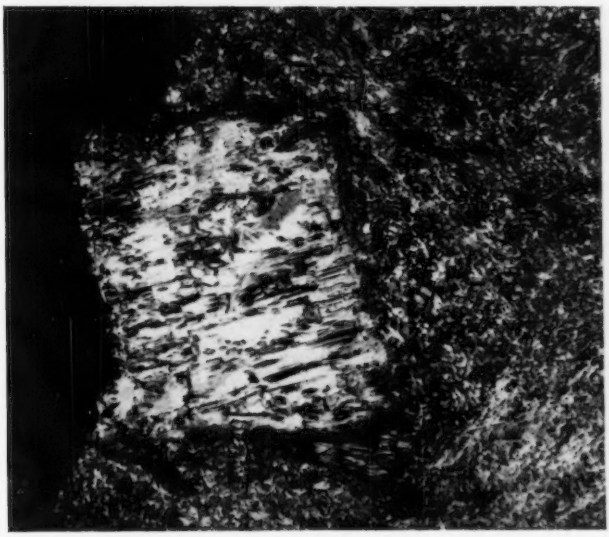


FIG. 2.

PHENOCRYSTS OF CHEQUER-ALBITE IN KERATOPHYRE, RANGIAWHIA PENINSULA. CROSSED NICOLS, × 48. ONLY A FEW DISORIENTATED FELDSPAR LATHS ARE ENCLOSED.

FIG. 1.

POTASH FELDSPAR (DARKER GREY) REPLACING ALBITE PHENOCRYST IN KERATOPHYRE, GREAT ISLAND, THREE KINGS GROUP. CROSSED NICOLS, × 45.

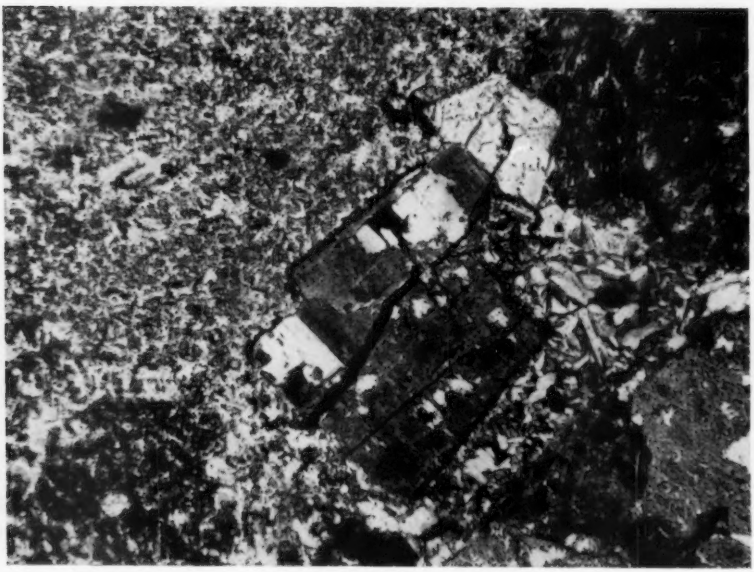


FIG. 2.

POTASH FELDSPAR (GREY) REPLACING ALBITE (WHITE) IN KERATOPHYRE, GREAT ISLAND. CROSSED NICOLS, × 45.

In microscopic texture the rocks fall readily into the groups proposed by Reed (1895) in his excellent descriptions of Welsh Ordovician acid lavas, namely the *microlitic*, the *cryptocrystalline* (for which the term *felsitic* is to be preferred), and the *micropoikilitic*, with the addition of a class of *radiolitic* [1] and *spherulitic* types. The felsitic and radiolitic-spherulitic types often retain relict glassy structures in great perfection, outlined by finely-divided chlorite or biotite. The relict textures of all the types show strong similarities to those of modern rhyolites. In particular, some of the rocks resemble closely the ignimbrites (or welded rhyolitic tuffs) of younger volcanic areas. It is of interest that this last analogy was drawn long ago by Dakyns and Greenly (1905) in respect of felsitic rocks from the Snowdon rhyolite assemblage of Wales, and again, recently, by Amstutz (1950) in describing quartz-porphyries associated with spilites in Canton Glarus, Switzerland.

The Alkali Feldspars

Albite.—In the northern New Zealand rocks the albite invariably has the optical properties of the low temperature form. The albite phenocrysts may be divided into two types, those with albite lamellation continuous across the crystals, which are often mottled with sericite and chlorite inclusions, and those with incipient or well-developed chequer-structure, which are usually of water-clear albite. The chequer-albite is twinned on the Albite law, but with short lamellae which, after continuing for a length a little greater than their width, either wedge out or are abruptly truncated by planes parallel to (001): such short lamellae, in the same orientation but offset from one another, occupy the whole area of the phenocryst and, in sections normal to (010) when one part of the twin is in extinction, produce a chequered pattern. The crystals are tabular parallel to (001) and often show a simple Manebach twin. Chequer-structure is found in all textural types but the micropoikilitic rocks are remarkable in that their phenocrysts are almost invariably chequered.

This feature of the albite of keratophyres has been recorded from many occurrences (e.g. Beskow, 1929; Cox, 1915, p. 328; Flett, 1911, p. 93, and 1913, p. 58; Götz, 1937; Lehmann, 1941). In the Kiruna "syenite-porphyries" Geijer (1910, pp. 63, 127) noted that the "striped" (= chequered) structure is characteristic of very pure albite (p. 63); sometimes the crystals are perthitic, but there is a class of phenocrysts with "striped" twinning and a simple Manebach twin, which are without perthitic lamellae (p. 127). In conformity with this it has been noticed in the New Zealand keratophyres that the chequer-albites often lack intergrown potash feldspar, even in potassic rocks,

[1] Loewinson-Lessing's term: see Johannsen (1939), pp. 230, 231.

while such intergrowths are of common occurrence in the evenly-twinned type.

Chequer-structure has so often been described as associated with secondary albite that some writers (e.g. Gilluly, 1933; Götz, 1937) have used it as supporting evidence of albitization. On the other hand, in plutonic rocks it has sometimes been regarded as characteristic of late magmatic crystallization in the presence of abundant volatiles (Tilley, 1919). The process of progressive recrystallization and modification of feldspars with slow cooling for plutonics does much to harmonize these two interpretations.

In the New Zealand rocks several stages in the development of the structure are to be found. The earliest stage is apparently represented by phenocrysts of albite with the most indefinite ghostly outlines, which show internally a kind of shadowy extinction resulting from their " composite " structure. Each is constituted of a mosaic of grains in slightly different orientations, the effect being that of a single crystal being formed by the welding together of independent laths and continuing to incorporate groundmass laths about its margins (Plate V, fig. 1). This feature may be compared with that described by Flett (1911, p. 93) from the pink intrusive " porphyry " associated with pillow lavas in Tayvallich Peninsula. Examples can be found ranging from such vaguely-bounded " composite " crystals to those in which a higher degree of ordering of the individual laths prevails. In this case the laths are arranged in three tiers like unevenly laid tiles on a roof. Those of the layer parallel to the side of the triangular crystal are largely united into homogeneity, the next two layers have their laths inclined to each other in a very obtuse chevron pattern, and the apex of the triangle, separated from the rest by a row of very variously orientated pieces, is fairly homogeneous but contains irregularly disposed crystal flakes. The boundaries between the tiers are normal to the length of the laths (which in turn is parallel to the albite lamellation) and roughly conform with the Manebach twinning often seen in the chequered crystals. The margins of the crystal just described are quite sharp and straight, in spite of its xenomorphism.

Finally, chequered crystals of mainly regular structure can be found, with just a few disorientated laths within them (Plate V, fig. 2).

On these grounds, the most plausible suggestion is that chequer-albite is here due to porphyroblastic growth proceeding from numerous centres and approaching axial parallelism by successive recrystallizations. The sequence traced is complicated by the fact that some chequered phenocrysts are being replaced marginally and internally by potash feldspar in fine crystallization, so that all vaguely-bounded crystals cannot be regarded as growth stages of chequer-albite. Moreover it is not necessary to suppose that all these phenocrysts grew *de*

novo in the rock. It seems more likely, indeed, that partially formed phenocrysts separated in the original rock and that these have been rebuilt and completed as albite phenocrysts. Nevertheless some of these seem to have formed from very vague and imperfect beginnings.

The mobility of albite in these rocks is supported by cases of healing of fractured albites by pellucid new growth of the same mineral, by the occasional presence of quartz-albite veinlets on a microscopic scale, and by the presence of albite in amygdales.

Potash Feldspar.—The phenocrysts of the keratophyres are remarkable also for the stages they show of replacement of albite by orthoclase-cryptoperthite (usage of Tuttle, 1952a) of variable optical properties. A complete series of examples can be found in the New Zealand rocks, from the earliest beginnings of potash feldspar development in the albite, and replacement around the margins and along cracks, to almost complete, and finally complete elimination of the islands of albite remaining in the potash feldspar (Plate VI). The poorly refringent and birefringent potash feldspar is often clearly distinguishable from the albite, but staining with sodium cobaltinitrite (following the procedure of Chayes, 1952) has been found very useful in determining the amount of potash feldspar in the phenocrysts and is virtually indispensable for differentiating it in the groundmass.

The phenocrysts in any rock slice are replaced to varying degrees and the amount of replacement in the rock as a whole may vary within one lava-flow. For example the upper keratophyre on Great Island, Three Kings Group, in several specimens (4609, 4610, etc.[1]) shows only incipient replacement of phenocrysts with small maculae of potash feldspar in the albite, though the groundmass shows a definite stain, whilst in others (6208) replacement of phenocrysts is well advanced and the groundmass takes a deep stain.

The frequent intergrowth of albite and potash feldspar was observed early in the history of keratophyre study, as may be seen from the descriptions by Lossen quoted by Teall (1888, p. 370). So far as the writer is aware it has, in the past, been described as perthitic and ascribed to exsolution either explicitly (Götz, 1937, pp. 182, 183) or by implication (Beskow, 1929, p. 93; Geijer, 1910, pp. 2–4, 25, 45, 127, and 1931, pp. 192, 218; Lehmann, 1949, p. 10; Sundius, 1915, p. 231).

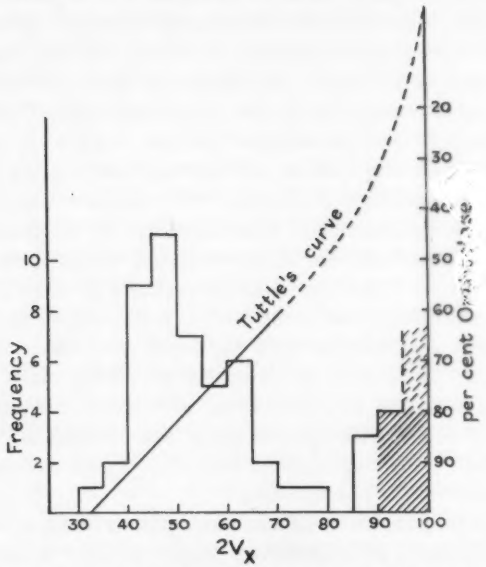
The following points oppose this view.

(1) The irregular nature of the intergrowth, the marginal encroachment of potash feldspar around the borders of albite crystals and its advance along cleavages and other cracks. This may progress until only little islands of albite remain, enveloped in potash feldspar, and

[1] Numbers refer to the catalogue of the Geology Dept., Auckland Museum, New Zealand. Duplicate material is lodged in the Museum of the Dept. of Mineralogy and Petrology, Cambridge University.

finally replacement becomes complete. The bounding of albite remnants by the cleavages is noteworthy because the perthitic lamellae of exsolution perthites are seldom parallel to the (001) and (010) cleavages. X-ray studies have shown good reason why they are parallel to (100) (Bragg, 1937, p. 241). On the other hand, it is clear that replacement will readily conform with the cleavage pattern.

(2) The variable proportions of the constituents of the intergrowth in different crystals in the same slice. Various stages in the process of replacement can be seen in close association. Exsolution, on the other hand, would imply a relatively small range of variation in the ratio of potash to soda feldspar, appropriate to the homogeneous precipitate



TEXT-FIG. 1.—Frequency distribution of values of the optic axial angle in replacement orthoclase. The cross-hatched area represents a range within which crystals retain Albite twinning. Tuttle's (1952*a*) curve connects $2V_x$ with composition.

at the temperature of crystallization of the phenocrysts. The continuity in the stages of orthoclase replacement is important also in view of the emphasis that has been laid on the replacement of potash feldspar by albite in other occurrences. The direction of the process *from* albite *to* potash feldspar is held to be securely established by this continuity, in combination with the textural evidence cited in (1), and that in (3) below.

(3) The increase in orthoclase in the groundmass, forming parts of otherwise albitic microlites, concurrently with the increase in potash

feldspar in the phenocrysts. More significantly, potash feldspar may at the same time develop from former glass, which still retains its perlitic cracks.

(4) The variation in alkali ratio, often with antipathetic changes in soda and potash, in different parts of the same lava flow.

(5) The production of rocks showing these features thought to indicate replacement, which have so high a content of potash that they could not have crystallized from a magma derived along the liquid line of descent from a more basic parent melt.

In replacing albite, the *a*-axis of the orthoclase coincides with that of the albite, and the (001) cleavage can sometimes be seen to continue from one mineral to the other. The potash feldspar is optically monoclinic and has the optic axial plane normal to (010). The optic axial angle ranges from 34° to 80° about X (with a few values between 86° and 90°) in different crystals and in different parts of the same crystal. Text-fig. 1 shows the frequency distribution of 54 measured values of the optic axial angle. In any one grain the lower values of 2V were read on clearer portions of the grain, the larger on finely mottled or cloudy material. This cloudiness is not due to visible inclusions, and is probably caused by the higher proportion of cryptoperthitic albite in parts with higher 2V.

The extinction angle X: *a* in sections normal to Z is from 5° to 9°, but since such sections cannot be used for measurement of 2V, except when this is very large, no correlation could be established between optic axial angle and extinction angle. One crystal with $2V_x$ 89° showed X: *a* 16° which represents a drop from the value of 19° to 20° for albite.

Representative values of the refractive indices (slice 4428) are: *n*X $1{\cdot}522 \pm 0{\cdot}001$, *n*Y $1{\cdot}527_5 \pm 0{\cdot}0005$, *n*Z $1{\cdot}529_5 \pm 0{\cdot}001$. The values for $2V_x$ in this slice averaged 43° for clear parts of crystals, 53° for cloudy parts, and had a range from 38° to 75°. These refractive indices and optic axial angles may be compared with the data of Fenner (1936) from Yellowstone National Park given below:—

	Phenocrysts.	*Metasomatized groundmass material.*	*Pure potash orthoclase (in veins).*
	Or 50·5 Ab 44·3 An 5·2	Or 60·27 Ab 39·73	
*n*X	$1{\cdot}523_3$	1·522–1·524	1·518
*n*Y	$1{\cdot}528_4$	1·528	1·523
*n*Z	$1{\cdot}529_8$	1·529–1·530	1·524
$2V_x$	35–42° r > v	33–52°	—

An X-ray oscillation photograph of part of a phenocryst without visible albite remnants showed the presence of both soda and potash feldspar, confirming the view that the material replacing the phenocrysts is cryptoperthitic and supporting the inference that the variation in its optical properties depends upon the amount of unreplaced albite present. It was not possible, however, to obtain material of uniform optical properties suitable for composition determinations by X-ray methods.

A powder photograph was taken of potash feldspar replacing original glass, in a rock with replaced phenocrysts, and this proved, as was expected, to be *pure* potash feldspar, as indicated by the spacing of the $\bar{2}01$ reflexion (Bowen and Tuttle, 1950, p. 491). The lack of separation of the 130 and $1\bar{3}0$ reflexions (Mackenzie, 1952a) indicated monoclinic lattice symmetry in this latter material. Comparison of its optics with that of the material replacing phenocrysts was not possible as it was of spherulitic crystallization.

The replacing potash-soda feldspar extends over much of Tuttle's (1952a) orthoclase-cryptoperthite series, but the continuity of the replacement series is not well demonstrated over the whole range of compositions. Although the appearance of the potash-soda feldspar varies in different parts of the same crystal, from clear material with low 2V (more potassic) to cloudy material of larger 2V (more sodic), there is always a sharp change in refractive index and birefringence at the boundary between potash-soda feldspar and albite. This lack of complete optical gradation between the two types is shown also by the frequency distribution of observed values of the optic axial angle (Text-fig. 1) in which there is an absence of values of $2V_x$ between 81° and 85°. Material with $2V_x$ greater than 90° shows Albite twinning.

These facts suggest that there is a discontinuity in the series from albite to orthoclase, or that the initial change from albite to potash-soda feldspar takes place over a narrow zone of rapid transition. A discontinuity was supposed by Spencer (1937, p. 490) to exist between Or_{20-30} and Or_{40-50}, while Oftedahl (1948, p. 67) inferred a break from Or_0 to Or_{30}. Although Tuttle (1952a, p. 564) states that the optics of the series appear to be gradational, his curves of $2V_x$ and refractive indices against composition lack support from Or_{40} to Or_3.

Elsewhere Tuttle (1952b) has shown that analysed alkali feldspars from extrusive and hypabyssal rocks cover the whole range of composition from Or_0 to Or_{75}, while those from plutonic rocks fall into two groups with maximum frequencies of Or_{12} and Or_{72} separated by a gap. The existence of this gap in the compositions of plutonic alkali feldspars is ascribed to unmixing of primary feldspars of intermediate composition by stages, producing successively cryptoperthites, perthites, and separate crystals of potash-rich and soda-rich alkali feldspar, the

composition gap being located around the summit of the solvus curve of the alkali feldspar system.

In the case of the keratophyres, on the other hand, the cryptoperthite series results from replacement of albite by potash feldspar, and it would seem that the feldspar composition must have crossed the gap between albite and material with $2V_x \leqq 80°$. The absence of a gradational change over this range suggests that there is a real, though slight, break in the orthoclase-cryptoperthite series in conformity with Spencer's and Oftedahl's views, and that this break is of a structural nature as Spencer (1937, p. 490) implied. In the high temperature alkali feldspar series Donnay and Donnay (1952) found that the symmetry change from triclinic sodic to monoclinic potassic feldspar takes place at Or_{33}, although solid solution may be considered complete for all practical purposes. It is of interest to note that the position of the break inferred by Spencer in the orthoclase-cryptoperthite series is at about this point also (Or_{35}), while it is from this composition that Mountain's series of types richer in lime, varying continuously from $Or_{35}Ab_{65}$ to $Ab_{65}An_{35}$ diverges from the alkali feldspar series (Mountain, 1925).

At the point of symmetry change between triclinic albite and monoclinic, or pseudo-monoclinic, orthoclase-cryptoperthite lattice distortion would be at its maximum. In the presence of sufficient lime this might be expected to be relieved by the entry of the intermediate-sized calcium ion to produce Mountain's series. In the case of replacement phenomena where potassium is replacing sodium, the lattice instability might induce the rapid entry of additional potassium ions replacing sodium over a range of composition about the point of symmetry change, so that optically and chemically intermediate material is seldom observed.

Comparisons with keratophyres and old rhyolites from other areas show that the features held to indicate replacement of albite by potash feldspar are widely developed.

In the rocks of the classic Kiruna district the same relations between the feldspars obtain as in the New Zealand rocks, from the presence of maculae of potash feldspar (often bounded by the cleavages) in the albite to little rectangular remnants of albite in potash feldspar. That the same variation from place to place is found as in New Zealand may be inferred from the original descriptions, in which the feldspar is referred to as microperthite in which the potash feldspar component is subordinate and sometimes absent (Geijer, 1910, pp. 3–4, 45, 127, etc.; also 1931, p. 192). Sundius (1915, p. 231) describes them as primary quartz-microperthite-bearing rocks and divides them into two groups, one with important potash feldspar, the other with albite only. He thought that the two kinds might have different geological relations,

though Geijer (1910, pp. 14–19) had carefully demonstrated that on Kirunavaara they pass into one another without a break. The controversy is never satisfactorily resolved. Sundius admits that both potash-rich and albite-rich types occur together and have not been separated during the field work (1915, p. 235). He envisages the possibility of albite replacing potash feldspar, but not the reverse, in his statement (p. 236) that we do not know how much of the plagioclase is original " since we cannot exclude the possibility that a displacement of potash feldspar has occurred, though this is in no case directly demonstrable ".

From study of the New Zealand rocks, the confusion that arose over the potash feldspar-bearing and albitic types can be seen to have a possible source in the capricious nature of the process of potash enrichment.

Amongst the German keratophyres, Lossen's descriptions of examples from the Harz Mountains (quoted in Teall, 1888, p. 370) show the feldspar relations to be the same as those in the New Zealand rocks and Teall regards the features of the feldspar as diagnostic. The keratophyres from the Lahn region of Hesse have been closely studied in recent years and a large series of analyses of them published (Götz, 1937; Lehmann, 1941, 1949). The wide variability of the alkali ratio is the most striking feature of these analyses and they leave no doubt as to the completeness of the gradation from highly sodic to highly potassic types. Götz (1937, p. 182 sqq.) has given detailed descriptions of the keratophyres. He stresses the variability of the relations between the alkali feldspars and the wide occurrence of microperthitic types which are the characteristic feldspars of the Lahn rocks. He states that on the whole, amongst the phenocrysts, a perthitic, and often coarsely perthitic type predominates, but at the same time there are abundant " moirée-like " cryptoperthitic crystals and completely homogeneous, uniformly-extinguishing sections. Pure orthoclase is rare amongst the phenocrysts, soda-orthoclase that has not exsolved being more frequent, whilst yet more frequently are found crystals of pure albite in one part of the rock and pure anorthoclase in another. In the groundmass orthoclase is, as a rule, more important, but microperthite, soda-orthoclase, and anorthoclase are not absent.

In his view crypto- or microperthite may be chiefly the product of unmixing of potash-soda feldspar, especially since regular, if vaguely-bounded, albite inclusions are to be found. He remarks that the albite areas in the microperthite never have the usual spindle form, but are roughly isometric and are commonly bounded by the cleavages. In general, intercalations of potash feldspar are more frequently identified than albite in potash or potash-soda feldspar.

Götz's descriptions make abundantly clear the similarity of many features of the Lahn keratophyres to the New Zealand types. The

present writer has had the opportunity of examining a suite of Lahn keratophyres in the Museum of Petrology, Cambridge, and in these the optical properties of the alkali feldspar conform with those of the replacement orthoclase of the New Zealand rocks. Examples with anorthoclase form and twinning were not observed. In many of the Lahn keratophyres examined, some of which were stained, the features taken as indicating replacement of albite by potash feldspar are prominent. The clear marginal relationship of potash feldspar to glomeroporphyritic groups of albite crystals joined by sutured boundaries shows that replacement took place after the phenocrysts had been swept together and sutured. In stained slices the amount of potash feldspar varies considerably from crystal to crystal, as Götz pointed out in his descriptions. These facts indicate that we have here another case of part of the albite being progressively replaced by potash feldspar, though there is also evidence of the late crystallization of albite (with chequer-structure) in these rocks.

Amongst British rocks replacement texture has been noted in rhyolite tuff from Halfway House, Snowdon, and the preponderance of potash over soda in some of the rocks analysed by Williams (1927) could not have been attained along the liquid line of descent. Cox (1915, pp. 312, 317–8) records phenocrysts of soda-orthoclase along with albite in Pembrokeshire keratophyres. Flett (1911, p. 93) in describing porphyry associated with spilites at Tayvallich mentions Becke's belief that chequer-albite is due to replacement of orthoclase by albite and continues: "There is no proof of this in slides of the porphyry, but it may well be the case. Sometimes one-half of a Carlsbad twin is checker-albite, the other half microperthite; in other cases the checker-albite forms broader or narrower patches in a matrix of orthoclase, making a coarse and abnormal microperthite."

Beskow (1929, p. 83) finds the feldspar in the keratophyres of Södra Storfjället to be mainly albite, but exceptional types have potash feldspar as coarse perthitic drops or patches in the albite.

A number of cases are on record of a similar replacement in intermediate or basic rocks, where, however, it is apparently a rarity (e.g. Sargent, 1917, Tomkeieff, 1941, Roever, 1942, pp. 225, 275).

Features of the Groundmass

The process of potash enrichment has also affected the albite crystals of the groundmass. Staining shows, however, that the greatest effect has been upon formerly glassy material. Microlitic keratophyres, with little original glass, are sodic in composition and replacement of albite laths by potash feldspar is usually only slight. This type is, however, commonly associated with micropoikilitic types, as was observed by Reed (1895, pp. 164–5), who ascribed it to the devitrifica-

tion of glassy patches (now micropoikilitic) in rocks in which crystallization was just beginning. In northern New Zealand microlitic types form the margins of flows the interiors of which are micropoikilitic.

In felsitic types, owing to the fineness of grain, the distribution of the stained potash feldspar is difficult to determine. The quartz-albite-orthoclase mosaic is often finer around groups of phenocrysts and staining shows a higher proportion of potash feldspar here. Lines of yellow stain also tend to follow the curving flow lines marked out by wisps and streaks of chlorite and sometimes pumiceous shreds are picked out.

The micropoikilitic type of groundmass stains deeply, and it is clear that in these rocks it is not composed of quartz enclosing feldspar microlites, as described by Reed (1895, p. 166) and Harker (1935, p. 153). The albite laths are embedded in roughly circular areas of vaguely radiating material with a mottled appearance, birefringence somewhat less than that of quartz and a refractive index generally a little below that of Canada balsam, which has rather shadowy extinction reflecting its radiating structure. About the borders of the micropoikilitic plates lie little irregular pools of clear quartz, in discontinuous rings squeezed in between opposing cryptographic areas. The micropoikilitic plates themselves take an even cobaltinitrite stain. Chemical analysis shows that they are composed of a cryptographic intergrowth of quartz and potash feldspar, and in the more coarsely crystallized types the graphic intergrowth sometimes becomes microscopically visible.

An analysis of a micropoikilitic rock gave the calculated composition (ignoring 7 per cent of minor constituents and recalculating to 100): quartz 42 wt. per cent, orthoclase 45 wt. per cent, albite 13 wt. per cent. Micrometric analysis showed only 0·8 wt. per cent albite phenocrysts, 4 wt. per cent interstitial quartz, 93 per cent micropoikilitic material, and 2 per cent other constituents. The groundmass, then, has practically the same composition as the rock as a whole, and the cryptographic part must be composed of 46 wt. per cent quartz and 54 wt. per cent orthoclase. The modal quartz and cryptographic material together contain 48 wt. per cent SiO_2 and 52 wt. per cent potash feldspar, thus lying almost exactly on the silica-potash feldspar eutectic as determined by Schairer and Bowen (1947).

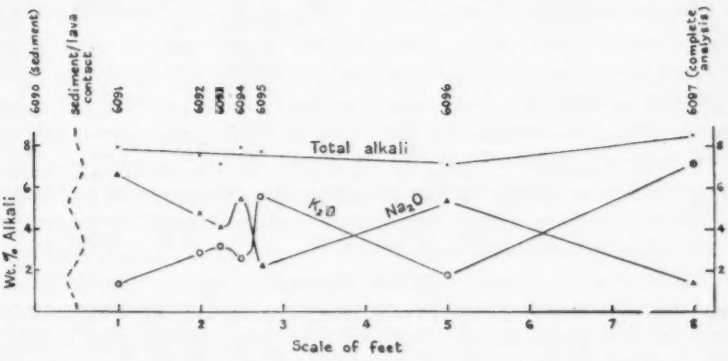
Chemical Composition and Petrogenesis of the Keratophyres

Eight complete analyses are available of keratophyres from northern New Zealand, with determinations of alkali content of a further six rocks. These data are presented in Table I.

The most striking feature of the analyses is the variability of the

alkali ratio. In order to test the range of variation in a single flow, the rapidity of the changes from point to point, and any possible relationship to the flow margin, the series of alkali determinations was made on specimens from across the lower part of a keratophyre flow, which showed a change in texture from microlitic to micropoikilitic from the margin inwards. The results, presented graphically in Text-fig. 2, show that the alkali ratio varies widely, irregularly, and very rapidly. Total alkali content, however, remains roughly the same throughout, at a value of 8 per cent and in this the rocks conform with the great majority of rhyolites.

All the analysed keratophyres have normative quartz + orthoclase + albite in excess of 80 per cent, so that their compositions may be



TEXT-FIG. 2.—Alkali variation at the lower margin of a keratophyre flow, Rangiawhia Peninsula, North Auckland.

represented on Bowen's (1937) diagram of the system carnegieite-kaliophilite-silica, the appropriate portion of which is reproduced in Text-fig. 3. The compositions of rocks produced by pure fractional crystallization should fall in the trough of low melting temperatures indicated on this diagram, towards which, as Bowen has shown, the residual liquids of this process are directed. He used the composition of Daly's average of 102 rhyolites (" G " of Text-fig. 3) to indicate the conformity of natural rhyolites with requirements of the theory of fractional crystallization. As another representation of this residual composition, the average of 8 analyses of rhyolitic glasses from Glass Mountain, Northern California (Anderson, 1933) believed to have been erupted only about 500 years ago, and therefore probably immune from secondary changes, has been plotted as the point " A " in Text-fig. 3. This has been done because a survey of published analyses shows a marked variability in the alkali ratios of rhyolites. In a statistical study

TABLE 1

Chemical Analyses

	6249	5299	6375A	6097	B (cf. 4609)	4427	4720	C
SiO_2	70·26	71·23	71·34	74·80	75·10	76·07	76·08	76·81
TiO_2	0·68	0·47	0·36	0·01	0·22	0·14	0·15	0·19
Al_2O_3	12·58	13·74	13·66	13·59	12·84	12·50	12·02	11·70
Fe_2O_3	3·30	0·45	3·28	1·05	0·70	0·75	1·05	0·35
FeO	2·00	1·18	1·48	0·29	1·36	0·89	1·09	0·39
MnO	tr	—	tr	—	0·04	—	—	tr
MgO	0·36	0·27	0·34	0·24	0·30	0·40	0·50	0·30
CaO	1·26	0·58	1·26	0·25	0·32	0·13	0·18	0·30
Na_2O	4·77	0·53	5·74	1·42	5·12	4·37	1·84	2·73
K_2O	2·88	10·31	0·87	7·10	2·39	3·31	4·99	5·45
H_2O^+	0·97	0·58	1·28	0·97	0·95	1·08	1·44	0·96
H_2O^-	0·27	0·06	0·41	0·33	0·27	0·22	0·31	0·51
P_2O_5	0·52	0·13	0·05	0·08	0·04	0·03	0·02	0·08
CO_2	—	—	—	—	0·03	—	—	0·02
ZrO_2	—	—	—	—	0·01	—	—	0·01
S	—	—	—	—	0·06	—	—	0·02
Cl	—	—	—	—	0·02	—	—	tr
NiO	—	—	—	—	—	—	—	0·01
BaO	—	—	—	—	0·05	—	—	0·05
SrO	—	—	—	—	tr	—	—	0·01
Sum	99·85	99·53	100·07	100·13	99·82	99·89	99·67	99·93
D	2·68	2·60	2·65	2·59		2·61	2·58	

C.I.P.W. Norms

qu	29·64	26·64	31·32	38·88	34·28	37·14	45·12	38·88
or	17·24	61·16	5·00	42·26	14·14	19·46	29·47	32·25
ab	40·35	4·72	48·73	12·05	43·31	36·68	15·72	23·06
an	2·78	1·95	6·39	0·28	1·59	0·56	0·83	1·39
cor	0·61	0·92	0·92	3·37	1·25	1·63	2·65	0·82
hy $MgSiO_3$	0·90	1·06	0·90	0·60	2·43	1·00	0·92	1·06
hy $FeSiO_3$	—	0·70	—	—		0·66	1·30	—
mt	4·41	0·70	3·48	0·93	0·91	1·16	1·62	0·46
ilm	1·37	0·91	0·76	—	0·43	0·30	0·30	0·30
ap	1·34	0·34	—	0·34	—	—	—	—
hm	0·32	—	1·12	0·48	pr.	0·12	—	— pr. 0·15
water	1·24	0·64	1·69	1·30	1·22	1·30	1·75	1·47
Sum	100·20	99·74	100·31	100·49	99·68	99·92	99·68	99·84

Partial Analyses

	6091	6092	6093	6094	6095	6096
Na_2O	6·59	4·67	4·03	5·44	2·18	5·39
K_2O	1·34	2·78	3·08	2·56	5·55	1·80

Reference

6249 Dark-coloured keratophyre, summit of South West Island, Three Kings Group.

5299 Spherulitic keratophyre with relict perlitic structure, S.S.W. of Brodie's Creek.

6375A Dark green keratophyre, Mt. Camel.

6097 Light-coloured micropoikilitic keratophyre, 7 ft. 6 in. above base of flow, E. head of Brodie's Creek.
B (cf. 4609) Light-coloured felsitic keratophyre, near Castaways' Depot, Great Island (Bartrum, 1936).
4427 Keratophyre, Waiari, Rangiawhia Peninsula.
4720 Light coloured keratophyre with chloritic streaks, S. end Anaputa Cove, Rangiawhia Peninsula.
C "Albitic granophyre," Mt. Camel (Bartrum, 1929).
6091 Margin (probable base) of keratophyre flow, E. head Brodie's Creek, Rangiawhia Peninsula, 6 inches from contact.
6092 Ditto: 18 inches from contact.
6093 Ditto: 21 " " "
6094 Ditto: 24 " " "
6095 Ditto: 27 " " "
6096 Ditto: 4 ft. 6 in. from contact.
See also 6097 above.

Analyses B and C by F. T. Seelye. Others by M. H. Battey.

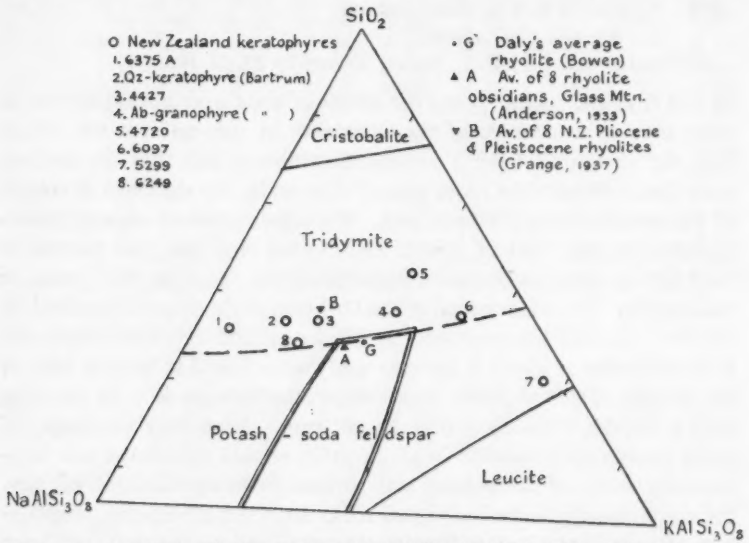
of 273 rhyolitic rocks,[1] using the excess of soda over potash or potash over soda as a measure of the variability of this ratio, it was found that the variations have a normal distribution and that the average rock has 0·80 per cent more potash than soda, the standard deviation of the sample being 2·90 per cent. The alkali ratios of a more homogeneous sample, that of British Ordovician rhyolites, are plotted in Text-fig. 4, along with some examples from more recent cycles of vulcanicity. The wide spread of the Ordovician rhyolites is matched by the New Zealand keratophyres. It will be noticed that total alkalis are fairly constant at about 8 per cent and this is found to be true also of the sample of 273 rhyolitic rocks where the average is 7·93 per cent with a standard deviation of 1·56 per cent. That Daly's average, of rocks presumably variable in alkali ratio, should fall in the low temperature trough of the residual melt surface therefore indicates, not that the compositions of the individual rocks analysed are due to solidification of a residual liquid of fractional crystallization, but that they have been formed from such a solidification product by later processes of redistribution of alkalis without overall gain or loss.

The New Zealand keratophyres do not fall in the low temperature trough but, with the exception of 4720 and 5299, lie in a line roughly parallel with the cotectic curve between silica and alkali feldspar, and within the silica field. The nature of this boundary is not precisely known, but Schairer (1950) states that the isotherms indicate a probable minimum on the cotectic curve. If this is so, and the minimum is not strongly displaced towards the silica-potash feldspar join, there is no possibility of producing a liquid composition such as 6097 by removal

[1] The sample was taken from Washington's Tables (superior analyses of fresh rocks) and was made up of 128 rhyolites; 27 liparites; 38 rhyolite-obsidians, -glasses, -pumices, liparite-obsidians, obsidians, marekanites and perlites; 17 pitchstones; 17 felsites, 17 keratophyres and quartz-keratophyres, 12 aporhyolites; 3 eorhyolites and rhyolite-porphyries, 4 soda-rhyolites, 4 tordrillites and 1 metarhyolite.

of albite-quartz mixtures in the form of microlitic marginal phases of the keratophyre, or otherwise. If the minimum were so displaced, the possibility of producing 6097 in this way would remain.

A highly potassic composition such as 5299 would, however, still demand a special explanation. Terzaghi (1935, 1948) has shown that the development of potash-rich rocks low in lime by fractional crystallization of liquids originally precipitating plagioclase is precluded by the phase relations of the feldspars. Schairer and Bowen (1947) showed that the boundary curve between the fields of orthoclase and plagioclase was a cotectic, and consequently rejected the hypothesis, preferred in



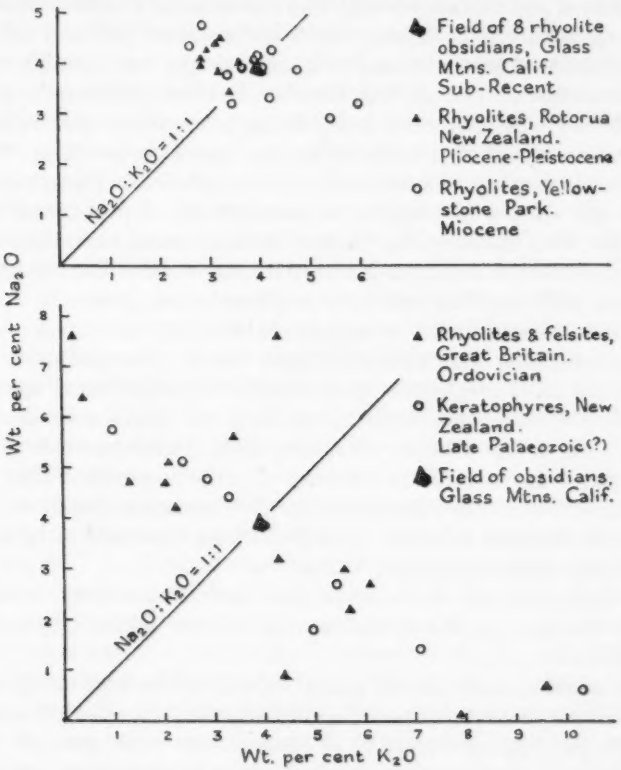
TEXT-FIG. 3.—Part of the system $NaAlSiO_4$-$KAlSiO_4$-SiO_2 (after Bowen, 1937) showing the low-temperature trough and the compositions of New Zealand keratophyres.

1928, of limited solid solution between orthoclase and anorthite permitting the development of potash-rich liquids by crystal fractionation. A rock of the composition of 5299 thus falls in an apparently non-magmatic field, and since the New Zealand keratophyres form a connected series this fact must be taken into account in considering their origin.[1]

[1] Higazy (1950) has collected analyses of potash-rich rhyolites that fall outside the low temperature trough, and some of which lie in the non-magmatic field. He regards metasomatism as probably important, but is inclined to consider that potash-rich magmas may be produced by immiscibility in silicate melts. If serious consideration is given to this second proposition in the face of the unfavourable experimental evidence, discussion of petrogenesis in terms of results from experimentally investigated silicate systems becomes rather pointless.

It must be assumed on available information that such a composition as "A" (Glass Mountain rhyolite glasses) or "B" (New Zealand Pliocene-Pleistocene hypohyaline rhyolites: Grange, 1937) represents the product of the liquid line of descent.

This being so, later changes must be invoked to account for the compositions actually met with. If it be assumed that, in a rock solidi-


Field of 8 rhyolite obsidians, Glass Mtns. Calif. Sub-Recent
Rhyolites, Rotorua New Zealand. Pliocene-Pleistocene
Rhyolites, Yellowstone Park. Miocene
Na2O : K2O = 1 : 1
Rhyolites & felsites, Great Britain. Ordovician
Keratophyres, New Zealand. Late Palaeozoic(?)
Field of obsidians, Glass Mtns. Calif.
Na2O : K2O = 1 : 1
Wt per cent Na2O
Wt. per cent K2O

TEXT-FIG. 4.—Scatter diagrams of alkali ratios of rhyolitic rocks of various ages. Data from Anderson (1933), Grange (1937), Iddings (1899), and Guppy (1931).

fying near the silica-feldspar cotectic of the residual system, later soda feldspar has volumetrically replaced potash feldspar, or *vice versa*, the resulting compositions will lie roughly along the boundary curve. On the basis of the chemical relationships of the rocks and the petrographic evidence of metasomatism, the New Zealand keratophyres are regarded as having been formed from a rock-type of a composition falling in the silica-rich end of the low-temperature trough, by enrichment in one or other of the alkalis and, to a less extent, in silica.

With regard to the sequence of metasomatism, it is clear from the petrographic evidence that in some cases the change involved has been first towards soda-enrichment (and possibly silification) and then towards enrichment in potash. Certain rocks (e.g. 4427) with a soda/potash ratio not far from unity show clearly by the way potash feldspar replaces their albite that their present composition is the result of potash enrichment and that earlier they were rich in soda. Potash-feldspar has even replaced chequer-albite, which is considered probably of lower temperature of formation than the normal type and possibly wholly porphyroblastic. This process, however, involves displacement of soda which may well have been precipitating concurrently elsewhere, and the occurrence of clear albite in vesicles, and with quartz in veinlets, shows late mobility of albitic and siliceous solutions. Though in most cases the veinlets in breccias are inconclusive of post-consolidation mobility, since the rocks may be flow-breccias, in one case albite-quartz veinlets cut a rock regarded as truly pyroclastic and albitic veinlets also traverse sediment close below keratophyre in one place.

In spite of long search no unequivocal evidence was found of albite crystallizing later than potash feldspar, but it is thought quite likely that it did so. The relationship of albite to orthoclase in the vesicles of radiolitic keratophyres suggests, but does not prove this. In other areas (e.g. Lahn district, Lehmann, 1941; Clarencetown district, Osborne, 1925, p. 115) the mantling of orthoclase with albite, albite veining and vesicle-fillings have led to the conclusion that albite is the principal replacing mineral, and this has long been held to be true of many pegmatites (Andersen, 1928).

It seems safe only to conclude that both alkalis were mobile at the same time, as indeed replacement of one feldspar by another demands.

The question of an external origin of the alkalis in the development of keratophyres must receive careful consideration, since the introduction of soda has been advocated by so many writers in the past. It seems clear from the relations described in the New Zealand rocks, and found to be general in the classical keratophyre occurrences, that both soda and potash enrichment must be considered together. It is important also to keep in mind the rather constant total alkali content of keratophyres and rhyolites, both old and young.

The suggested sources of soda, most of which have been proposed in connection with the "spilite problem" as a whole, include absorption of sea-water directly (Beskow, 1929, pp. 287–8) or from wet sediments (Daly, 1933, pp. 419–21), the chemical action of rising connate waters expressed from compacting sediments below, the incorporation of albite-quartz liquid melted out of geosynclinal greywackes at depth (Turner and Verhoogen, 1951, pp. 209–211) and

emanations from subjacent or intrusive trondhjemitic magmas (Gilluly, 1935). On the other hand, neither sea-water, not connate waters can be called upon to provide large amounts of potash in explanation of the associated potassic rocks. Since potash is differentially retained in the solids throughout the cycle of sedimentation and deposition, assimilation might be supposed to provide some solution, but there is no evidence of its occurrence, and the rapid variation in potash/soda ratio in a single lava flow puts severe limitations upon this speculation. If an external origin be proposed for the potash it must apparently be introduced in magmatically-derived solutions.

Parallel cases of potash-metasomatism due to magmatic emanations are most prominent in regions of andesitic eruption that lie either to one side of geosynclinal belts, or develop after folding, as for example in the western U.S.A., Transylvania, and the Hauraki Goldfield, New Zealand. In these areas potash metasomatism associated with propylitization is widespread, but irregular, and is ascribed, with a wealth of evidence, to watery solutions rising from below, presumably having their origin in a subjacent body of granitic rock. This is also true of the waves of metasomatism demonstrated by Fenner (1936) in the gravels and lavas of geyser basins at Yellowstone National Park, though in this instance the work of Allen and Day (1935) indicates that circulating ground waters, activated by magmatic emanations, play a large part in the hydrothermal activity.

In these cases, however, the metasomatic process is characterized by the clearly demonstrable passage of large volumes of solvent through open fissures. The channelways can be recognized and are margined by zones of intense metasomatism which diminish outwards in the country rocks as an " aureole " of metasomatism, as has been convincingly demonstrated by chemical analyses in the Hauraki Mining Area of New Zealand (Finlayson, 1909). The whole succession, lavas and sediments alike, traversed by the solutions, is extensively altered, though sediments are less susceptible than igneous rocks in many cases.

In the case of the keratophyres, the country rock is virtually unaffected, save for a slight hardening at contacts noted in New Zealand and in Pembrokeshire (Cox, 1915) and a few quartz-albite veins on a microscopic scale. Channels of access for solutions have not been observed in any of the occurrences described. Although in many of them (Kiruna, Södra Storfjället, Newton Abbot, Wales, Co. Wicklow, Lahn district, Dakota, Clarencetown-Paterson district, N.S.W., and New Zealand) the variability of the alkali ratio is most marked, the distribution of the variations is not readily explained in terms of centres of metasomatism.

From the information available at Yellowstone National Park,

Fenner (1936, p. 300) has calculated that the emanations from subjacent (granitic) magma contain the alkalis in the proportion of some 91 mol. per cent NaCl to 9 mol. per cent KCl, a result stated to be in accord with other data on magmatic sublimates. The potash so important in metasomatism is therefore thought (p. 301) to be acquired by carbonic acid leaching of the feldspar of the roof rocks of the intrusion. The leaching that Fenner envisages is non-selective of the alkalis (p. 299) which therefore enter the emanations in the proportions in which they are present in the rock being leached. This rock must therefore itself be potassic. If such leaching is required to charge the emanations with potash, additional assumptions must be introduced about the stratigraphic column below the level of observation, while the unaltered state of the rocks below the keratophyres becomes still harder to understand. These considerations show that great difficulties attend the conception of alteration of the keratophyres by emanations from below.

The redistribution of the alkalis within an individual lava flow, on the other hand, seems adequate to explain the compositions of the rocks, and receives strong support from the chemical data already presented, particularly the constancy of the total alkali content. In support of this conclusion may be quoted the view of Terzaghi (1948) that, though some of the potash of potash-rich rhyolites from the Esterel, France, may have been inroduced as gaseous or hydrothermal solutions from underlying magma, much of the enrichment is due to local leaching and deposition within the lavas themselves. The process is one which has been postulated also by Osborne (1925) to explain albitization and changes in alkali ratio along the strike in the Kuttung acid volcanics of New South Wales.

The actual processes in the zone of potash feldspar enrichment at Yellowstone Park seem, however, to be closely comparable with those in the keratophyre flows, and the selective precipitation of potash, demonstrated by Fenner to proceed to such an extent that the potash/soda ratio in the pore fluid is reduced to 1 : 34, goes far to explain the compositions of the potassic rocks. To his views may be added the belief that the partly glassy nature of the rocks contributes much to the differential metasomatism. Terzaghi (1935, p. 379) has pointed out that metasomatic potash-enrichment is common during the alteration of glassy lavas and it is of interest that Osborne (1925, p. 114) finds albitization does not affect pitchstones, but occurs only in non-vitreous types. The susceptibility of glass to potash enrichment has been noted in a number of cases in the New Zealand keratophyres while there is no evidence that it blots up soda to the same extent. The inhomogeneity of rhyolites with respect to their degree of crystallinity is well known, and these indications of differential binding of potash in glass, though

requiring the support of more numerous observations, may provide a partial explanation of the rapid variations in alkali ratio within the one rock body.

Conclusions

The keratophyres possess relict textural features (glassy flow structures, perlitic cracks, and pumiceous streaks) that link them closely with normal recent and late Tertiary rhyolites. These textures suggest very strongly that the rocks were exposed, in their initial eruption, to the same conditions as normal young rhyolites, that they were true superficial lava flows, were sometimes flow-brecciated, sometimes partly vesiculated to pumice and were rapidly cooled to glassy products.

On the other hand the mineralogy of the feldspars, the nature of the processes inferred for their precipitation, and the variations in bulk composition of the rocks show close parallels with the corresponding features of acid plutonic rocks. The mineralogy of the keratophyres thus implies that they were held for a long time at moderately elevated temperatures where recrystallization and internal migration of alkalis and silica were active processes, while the fact that perlitic glass has been altered to spherulitic potash feldspar shows that the rocks were already solid when this migration took place.

Their field relations show that they have been buried, probably deeply, and strongly folded.

There is no evidence of introduction of alkalis from without and their compositions are symmetrically distributed about the composition of a normal young rhyolite representing the lowest melting mixture of feldspars and silica.

The inference is plain that these keratophyres do not represent either the product of a special soda-rich magma, as so many writers since Dewey and Flett (1911) have supposed, or the result of wholesale introduction of alkalis as Gilluly (1935) claimed, but are simply rhyolites which have been buried to depths where the temperature was sufficiently high to allow recrystallization and limited mobility of alkalis and silica in pore fluids, so that the components were able to attain equilibrium under relatively low temperature conditions.

Sufficient has perhaps been said in comparing the mineralogical features of the New Zealand keratophyres with those from other areas to show that this conclusion, if valid, must apply to most keratophyres and old rhyolites.

Although keratophyres are not the product of any special magma, and do not form part of a separate igneous suite, they do represent a facies of mild metamorphism, and for this reason the retention of a separate name, marking them off from rhyolites of high-temperature mineralogy, is important. Keratophyre, in the original usage of

Gümbel and Lossen was employed for the rocks with which we are concerned and, in spite of Rosenbusch's application of it to rocks with sodic amphiboles and pyroxenes, it seems to have been retained in the original sense by most authors, though with various theoretical appendages. It is well entrenched in the literature in this sense and, with qualifiers indicating the preponderance of soda or potash when this is clearly marked, it is probably the most satisfactory term for rhyolites with a low-temperature mineral assemblage.

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The Geology of North-East Jan Mayen

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ABSTRACT

Field observations collected on a short expedition to Jan Mayen indicate that ankaramite was extruded at three different periods at the north-eastern end of the island, either as lava flows or pyroclastic deposits. Cycles of volcanic activity, in which the earlier lavas are ankaramite and the later ones trachybasaltic, have been recognized.

INTRODUCTION

THE island of Jan Mayen (71° N., 8½° W.) lies about 300 miles E. of of Greenland and at a slightly greater distance N.N.E. of Iceland (inset A to Text-fig. 1). It is about 35 miles long and elongated in a N.E.–S.W. direction with the volcanic cone of Beerenberg dominating the north-eastern end (inset B to Text-fig. 1). During a brief visit to the island in the summer of 1950 a rapid reconnaissance of this north-eastern end was carried out and 220 representative rock specimens were collected. These specimens are now housed in the Department of Geology at the University of Manchester and, since complete laboratory examination of them is likely to be delayed, it has been thought advisable to publish a separate account of the field investigations.

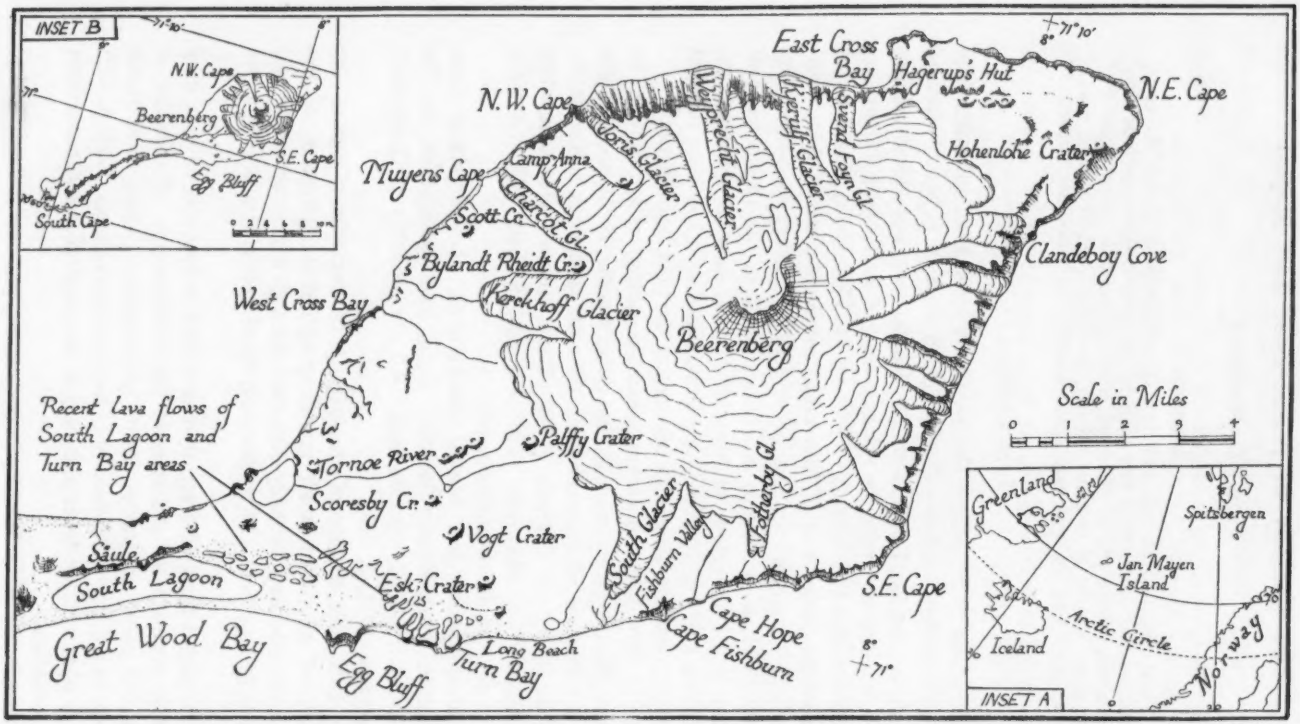
Prior to 1921 little was known about the geology of Jan Mayen. In that year Wordie gathered much field data (Wordie, 1926), and brought back a collection of approximately 60 rock specimens subsequently described by Tyrrell (Tyrrell, 1926). These authors refer adequately to the earlier work (Scoresby, 1820; Holmes, 1916, pp. 202–5; and the work of an Austrian expedition, 1882–3).

In north-eastern Jan Mayen trachybasalt of Beerenberg type, ankaramite, intermediates between them, and pyroclastics are the rocks most frequently encountered. All are probably of Quaternary age (Wordie, 1926, p. 745).

The distinctive feature of *trachybasalt* of Beerenberg type is the presence of abundant small glomeroporphyritic aggregates of bytownite-labradorite uniformly dispersed through a grey aphanitic and vesicular groundmass. Microphenocrysts of pyroxene and olivine are rare.

Large phenocrysts of green pyroxene and yellow olivine in a grey aphanitic groundmass characterize the *ankaramites*. Plagioclase is sparse and only very rarely forms phenocrysts. Thus, these rocks conform to the original definition of ankaramite (Lacroix, 1916, p. 182) but they often contain fewer phenocrysts than typical ankaramites from other localities.

Nomenclature of the Jan Mayen pyroclastics presents some difficulty.



TEXT-FIG. 1.—Main figure : North-East Jan Mayen showing localities referred to in the text. Map from surveys by the Austrian Polar Year Expedition, 1882–83, and the Imperial College Expedition,1938, with a few additions by the present author.
Inset A: The position of Jan Mayen relative to Greenland, Iceland, Spitsbergen, and Norway.
Inset B: Jan Mayen Island.

Although the size of the largest contained fragments is, in some cases, sufficient to justify the use of the term agglomerate, the tuffaceous matrix is usually so abundant that the term tuff-breccia is more appropriate. Finer-grained pyroclastics are termed lapilli-tuffs or tuffs (Wentworth and Williams, 1932, p. 52). Agglutinate is used in the sense defined by Tyrrell (Tyrrell, 1931, p. 66) and accepted by Wentworth and Williams (op. cit., p. 45).

Topography of North-Eastern Jan Mayen

North-east of a line drawn from West Cross Bay on the west coast to Turn Bay on the east a considerable part of the island is covered by a discontinuous ice-cap centred on the summit of the Beerenberg (Text-fig. 1). This ice-cap conceals the true nature of the volcanic pile which appears to have the form of a cone with concave sides, the slope of the upper surface increasing from 9° to 10° on the outer flanks to 30° to 40° near the summit (*ca.* 7,500 feet above sea level). Upwards it terminates in a crater approximately one mile in diameter and breached to the north by the Weyprecht glacier. Laterally it ends in coastal cliffs to the north, west, east, and south-east, but to the south-west passes beneath lavas of later date in the vicinity of Esk and Vogt craters and the Tornoe River. To the south-west, west, and north-east parasitic cones break the continuity of slope of the main cone—more may still be hidden beneath the ice for a number of dome-shaped, badly crevassed upheavals on the surface of the latter testify to pronounced local irregularities in the rock surface below.

In the extreme north-east, about North-East Cape, a low platform of lava extends seawards from the foot of the cliffs for distances of ½ to 1 mile. The cliffs here continue the line of the coast to the west and south so that these coastal lavas form a distinct fringe, which may be continued westwards as a foundation for the platform below the cliffs between Weyprecht and Joris glaciers. If so it is here completely hidden under morainic debris. More frequently narrow beaches of dark-coloured sand occur at the foot of the cliffs or else the latter rise almost sheer from the waves. Where accessible the cliffs afford good exposures of the volcanic rocks, and they have been studied in the Cape Fishburn area, and along the north and west coasts. Unfortunately the brevity of the visit prevented the work being continued along the east coast to complete the circuit of the ice-cap.

Observations in the Coastal Areas

(*a*) *Cape Fishburn.*—Wordie's record of thick deposits of tuff and agglomerate in this area was confirmed.

The lowest strata seen were at the base of the cliffs immediately S.W. of Cape Fishburn where the following succession was observed:—

Brown tuff with sparse scattered bombs of trachybasalt of Beerenberg type.	30 feet, at least.
Brown agglomerate with many fragments of ankaramitic material.	6–10 feet.
Ankaramitic debris containing boulders of red and black ankaramite and with an uneven upper surface.	1–3 feet.
Grey lapilli-tuff containing fragments of trachytic rock and other types not yet seen in situ at this north-eastern end of the island. The upper surface of this layer is very irregular and rises when traced south-westwards along the line of the cliffs so that the visible thickness of this deposit increases in this direction.	At least 4 feet (base hidden below shore debris).

Immediately south-west of the mouth of the Fishburn valley long scree slopes hide the solid rock up to a height of approximately 200 feet at which level dark-brown rudely stratified ankaramite agglomerate and elongated masses of ankaramite (up to 3 feet thick and several yards in horizontal extent) are exposed. One layer of ankaramite lying parallel to the stratification, is approximately 6 feet thick and has slightly scoriaceous surfaces. There is no evidence of veining of the adjacent agglomerate by this rock which is considered to represent a flow of ankaramite rather than a sill.

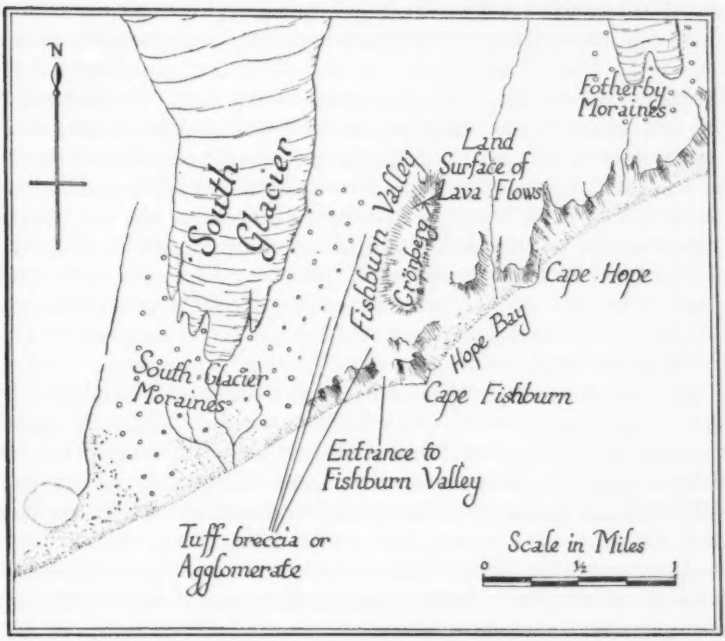
Tuff-breccia floors most of the Fishburn valley to the east of the moraines associated with the South glacier. At the head of the valley ankaramite is exposed at 1,050 feet and 1,120 feet and trachybasalt (of Beerenberg type), first encountered at 1,450 feet, is then continuously exposed up to 2,100 feet.

The cliffs behind Hope Bay are composed of agglomeratic material with at least eight layers of massive igneous material. On the hillside just above the cliffs a lenticular mass of ankaramite 20 yards long and 8 feet thick occurs in tuff-breccia. The slaggy upper and lower surfaces indicate its extrusive nature. Above the cliffs the land surface is an undulating platform of lavas, chiefly ankaramite and further exposures occur at a height of 1,030 feet above Cape Hope, and in the side of a valley to the west of the Fotherby glacier at a height of 1,280 feet.

Possibly the discontinuous nature of these ankaramitic masses prompted the earlier record of sills (Wordie, 1926) but in no example examined in 1950 could an intrusive origin be proved, and in some localities scoriaceous and slaggy surfaces clearly indicate that the ankaramite was extruded. Thus a period of extrusion of ankaramitic lavas can be recognized between that of agglomerate and tuff-breccia formation and the extrusion of the trachybasalts.

To the east of the Fishburn valley a ridge of tuff-breccia rising to a height of 800 feet runs inland from Cape Fishburn in a general N.N.W.–S.S.E. direction (the Grönberg, Text-fig. 2). At the northern end of the

ridge a rude stratification in this deposit strikes N. 51° E. and dips N.W. at 50°. Most of the rock fragments appear to be trachybasalt of Beerenberg type but occasionally other varieties of lava are also present, e.g. rare trachyandesite. The brownish ochreous matrix contains small fragments of crystals and in some layers crystals of augite and olivine are quite common. It is significant that these minerals are not those commonly forming phenocrysts in the Beerenberg trachybasalt, making it unlikely that the magma involved in the



TEXT-FIG. 2.—Sketch-map of the area around Fishburn Valley showing localities mentioned in the text.

explosive activity forming these tuff-breccias was of that type. On the ridge itself reddened and vesicular trachybasaltic lava forms discontinuous exposures up to 75 yards across and in one place the lava is completely underlain by tuff-breccia. Adjacent to these masses of lava the pyroclastic rock is dark brown in colour and very much more friable than usual, an effect attributed to slight thermal metamorphism of the tuff-breccia by the lava. Similar lava tops the high ground immediately above Cape Fishburn where it is much more continuous.

The ridge is not everywhere of the same height. A col separates the main Grönberg ridge from the high ground immediately above Cape

Fishburn and on it at least five thin (average width 2 to 3 feet) almost vertical dykes cut the tuff-breccia. Two trend N.–S., one veers between N.–S. and N.E.–S.W., and a fourth trends N. 35° E. The fifth is only poorly exposed. The rock of which they are composed is intermediate in character between Beerenberg trachybasalt and ankaramite. It is possible that these dykes served as feeders to the lava overlying the tuff-breccia of the ridge.

From a viewpoint above Cape Hope it is apparent that the mass of tuff-breccia forming the Grönberg ridge extends above the general level at which lava flows supersede tuffaceous and agglomeratic deposits. It is conceivable that the ridge was a buttress of pyroclastic rocks around which the later lavas flowed but the distinctive characters of the Grönberg tuff-breccia, viz., the presence of augite and olivine crystals, the disposition of the stratification, the occurrence of fragments of trachyandesite, etc., and the presence of broken fragments, not bombs, of trachybasalt, suggest that it is distinct from the earlier agglomeratic rocks in the sea-cliffs and formed during a phase of volcanic activity subsequent to the extrusion of at least some of the trachybasaltic lavas. Choking of a fissure on the site of the Grönberg ridge with tuff-breccia, possibly in part derived from pre-existing agglomeratic rocks, was followed by uprise of magma to form the dykes on the col and the lava overlying the tuff-breccia.

(*b*) *East Cross Bay—North-East Cape.*—Sections in agglomeratic rocks also occur in the cliffs at the north-eastern extremity of the island. Extensive screes hide the lower parts of these cliffs except at a few isolated localities, but exposures are good at higher levels. At first sight the cliffs appear to be stratified with layers of lava interbedded with red, yellow, and brown fragmental material. The friable nature of the latter makes the ascent of these cliffs a hazardous undertaking and although partial ascents were made the exploration of the cliff face was not as complete as could be desired.

Detailed inspection shows that no horizon can be followed far laterally and, despite considerable variation in grain size and nature of the fragmental deposits, no systematic changes could be discovered in vertical sequence or horizontal distribution. In the lower parts of the cliffs rudely stratified brown agglomeratic deposits contain highly vesicular fragments of fine-grained lava and fragments of pyroxene crystals, the presence of which suggests that the magma associated with the activity forming these pyroclastics was ankaramitic in nature. Corroborative evidence of this view is afforded by the occurrence of discontinuous layers of ankaramitic lava at heights of 130 feet and 200 feet in the cliffs immediately east of the Svend Foyn glacier. Not all the fragmental material in these cliffs is pyroclastic. Lenticles of dark-brown material occur which appear to be fossil soil or washes of

disintegrated basaltic debris and some of the reddish layers are bole containing occasional blocks of weathered lava. Near the top of the cliffs lavas predominate and agglomeratic material is quite subordinate, flows of phenocryst-poor ankaramite being interbedded with layers of reddened scoriaceous lava.

Dykes of ankaramite and olivine-bearing trachybasalt cut both the pyroclastics and the lavas in these cliffs. They split and re-unite in a rather complicated manner but in general strike E.–W. (variation from N. 72° E. to N. 105° E. was observed), dip very steeply northwards or are vertical and average about 2 to 3 feet in width. Marginal chilling of these dykes suggests that they are considerably later than the rocks which they cut.

About 2½ miles east of Hagerup's Hut the cliff line becomes less clearly marked, and the line of the steepest slopes swings round from the E.N.E.–W.S.W. direction to link up with N.–S. cliffs behind Clandeboy Cove (Text-fig. 1), leaving a low marginal platform up to a mile wide around the cape itself. The lava is a highly scoriaceous pyroxene-bearing trachybasaltic variety, much obscured by a mat of lichens, etc. Some of it, especially in the area about North-East Cape itself, has apparently flowed down on to the coastal platform from higher levels. About 3 miles east of Hagerup's Hut a low bluff, about 40 feet above sea-level, exposes a section through a channel eroded in vesicular trachybasaltic lavas and floored by 1 to 2 feet of a gravel-like deposit containing numerous fragments of the lava up to an inch across. The channel is now filled with ankaramite lava that elsewhere overlies the trachybasalts with apparent conformity but which in fact must be later by a period sufficient for the erosion of the channel and the deposition within it of a layer of gravel.

In the area around Hohenlohe crater (Text-fig. 1) the coastal lava platform is infested with small parasitic cones, mostly scoria cones but with lava flows in some of the larger ones, especially in the throat walls. In some cases the associated lava was ankaramitic, in some a trachybasalt carrying a few small phenocrysts of olivine and augite, but in most it was a reddened highly vesicular lava of doubtful type. Blocks of tuff-breccia have been found on or near the crater rim of some of the vents but tuff-breccia has not been seen in situ in any vent walls. The position of these tuff-breccia blocks on the surface of the cones suggests that they represent vent infillings subsequently broken up and blown out during a later phase of activity rather than fragments of earlier agglomeratic rocks blown out from below the lava platform during the initial opening of the vent. Volcanic bombs are not infrequent on the outer flanks of the larger cones. One is worth further mention. Of lava intermediate in character between trachybasalt and ankaramite, it encloses a sharply margined angular core of ankaramite, indicating

disruption of this latter rock type during the opening of the vent from which the bomb was expelled, and hence age relationships the reverse of those testified by the described exposures 3 miles east of Hagerup's Hut.

Vents from which lavas have been extruded late in the volcanic history of this north-eastern corner of Jan Mayen have been discovered at the top of the cliffs 1½ to 2 miles east of Hagerup's Hut. These vents form a line almost parallel to the cliff face and individual vents are elongated in the same general E.–W. direction. The extruded lava is very similar to the material of the dykes exposed in the cliff face, prompting the suggestion that members of this group of dykes may have acted as feeders to the vents. The lava from these vents flowed towards the cliffs, swamped over the edge, cascaded down to the beach below, and there deployed to form at least the western part of the coastal platform. In some localities, e.g. at the N.E. end of the line of E.N.E.–W.S.W. cliffs, falls of lava, frozen in place, have been preserved, phenocryst-poor ankaramite partly covering a succession of fine-grained lavas and bedded fragmental ankaramite at the top of an 800 ft. cliff. Here, then, we may recognize an earlier period of lava extrusion following the formation of tuffs and agglomerates and separated from a later one by a time interval sufficient for the cutting of sea-cliffs approximately 1,000 feet high.

(*c*) *North-West Coast.*—West of Svend Foyn glacier morainic debris obscures much of the solid rock between the three main glaciers of the north-west coast (Svend Foyn, Kjerulf, and Weyprecht). One outcrop of ankaramite occurs on the shore between the two last-named glaciers and exposures of fine-grained grey basalt underlie the ice at the centre of the snout of the Weyprecht, but otherwise no solid rock was seen along the shore between Svend Foyn glacier and the western moraines of Weyprecht glacier.

West of Weyprecht glacier a low bench separates cliffs trending E.N.E.–W.S.W. from the sea. The boulder-strewn surface of this bench falls westwards, being approximately 30 to 40 feet above sea-level near Weyprecht glacier and almost at sea-level near the moraines of Joris glacier. No rock in situ was seen on or in the low cliffs at the edge of this bench and it appears to be composed of morainic material. Scree covers the lower 300 feet of the cliffs behind this bench and above this level only lavas are exposed. The lowest exposures seen were of ankaramite but at a height of 550 feet lava exposures breaking through scree are of trachybasalt. At 650 feet ankaramite is again exposed, apparently forming a lava flow, but at 775 feet the lava is again trachybasalt of Beerenberg type and exposures of this rock continue to the top of the cliffs (*ca.* 900 feet). Especially in the upper part of the cliffs individual flows may possess reddened scoriaceous surfaces and may be

interbedded with reddish-brown layers of weathered lava. Similar interbedding of lavas (ankaramite) with reddened scoriaceous and weathered layers may be observed in exposures around the snout of Joris glacier, with the inference that extrusion was sub-aerial.

(*d*) *West Coast.*—Between North-West Cape and the snout of Charcot glacier rather inaccessible cliff sections appear from a distance to be made up of lava flows and the land surface above the cliffs as far south as the open valley west of the Kerckhoff glacier is composed entirely of lava flows where it is free from ice or morainic debris. Ankaramite and trachybasalt of Beerenberg type both occur. The trachybasalts appear to have flowed from the main Beerenberg vent but many of the ankaramite flows were extruded from parasitic vents on the flanks of the main cone.

Two of these parasitic vents, Bylandt Rheidt crater and Scott crater, were marked on the map produced by the Austrian Polar Year Expedition of 1882–83 but there are others, e.g. one near the ice margin between the Joris glacier and the Camp Anna ice tongue at a height of 2,280 feet, which were not marked, possibly because they have only recently been uncovered by the retreat of the ice. This last-mentioned vent is smeared with morainic material even at its summit, leaving no doubt that it was at one time covered with ice.

The frequency with which lavas from the parasitic vents flow seawards suggests that the land surface had its present westerly slope when they were poured out. Since, except for these ankaramite lavas, the land surface is largely of Beerenberg trachybasalt it is probable that the ankaramite flows from the parasitic vents are younger than the trachybasalts.

South of Scott crater a very rough land surface is cut in a flow of lava intermediate between ankaramite and Beerenberg trachybasalt, the source of which is apparently still hidden by the ice. A similar rock forms boulders in the moraines of Kerckhoff glacier and one such boulder carried angular inclusions of Beerenberg trachybasalt approximately an inch in diameter. If, as appears likely, some of these intermediate types are hybrids between the ankaramite and trachybasalt, this boulder indicates that solidified trachybasalt was assimilated by ankaramite, confirmatory evidence for the view that some, at least, of the ankaramites are younger than the trachybasalt in this region.

Pyroclastic rocks, rarely seen in this west coast area, are exposed at the base of the cliffs at Muyen's Cape. South of the Cape the cliffs are composed of a northerly dipping series of grey massive ankaramite flows interbedded with layers of reddened scoria similar to the series of ankaramite and scoria exposed near the Joris glacier. Northwards this lava series passes beneath an uneven heap of brown agglomeratic

material containing bombs of trachybasalt, which in turn is overlain by a series of disrupted trachybasaltic lavas. In the high cliffs of the Cape these trachybasalts end abruptly against an intrusion of ankaramite, to the north of which an agglutinate composed of reddened fragments of scoriaceous ankaramite is exposed. A fragment of this agglutinate caught up in and veined by the ankaramite leaves no doubt that its formation preceded the intrusion of the latter. Between the agglutinate and the morainic material associated with Charcot glacier a yellow lapilli-tuff is exposed. Fragments and crystals of pyroxene and olivine enclosed in the matrix of this rock indicate its affinities with ankaramite although contained lava fragments are usually too fine-grained for easy reference to any of the recognized types.

The agglutinate is similar to a cindery mass of scoriaceous ankaramite forming the vent-walls and rim of Esk crater to the south of Beerenberg. It probably represents a similar deposit on the site of a vent at Muyen's Cape which has been partially dissected by marine erosion, the ankaramite intrusion then being interpreted as a plug and the yellow lapilli-tuff as the pyroclastic deposit surrounding the vent. The nature of the vent, whether fissure or pipe-like, is uncertain, as inland from the Cape moraine associated with Charcot glacier hides the volcanic rocks.

The brown agglomeratic material and the overlying trachybasaltic lavas are probably earlier, since the former is quite distinct from the ankaramitic pyroclastics. Still earlier are the interbedded ankaramite and scoria layers.

Observations in Inland Areas

(*a*) *Central Mass of Beerenberg.*—Wordie's statement that the upper part of the Beerenberg is composed of trachybasalt lava can be confirmed. There are now outcrops of reddened ankaramite at 5,250 feet on the prominent south-western ridge of the mountain but below this level there are continuous exposures of trachybasalt flows through a vertical range of 1,000 feet. These lavas dip off the cone with the dip becoming steeper near the summit. Exposures are few on the flanks of the Beerenberg between 2,000 and 3,500 to 4,000 feet, at which level the surface of the ice slopes outwards at 9° to 10°. A long ridge of rock breaking through the ice cover behind Bylandt Rheidt crater is composed of almost non-vesicular Beerenberg trachybasalt, but elsewhere moraine covers most of the area free from ice.

(*b*) *South-West of Beerenberg.*—Below the ice margin (2,000 to 2,500 feet) to the south-west of Beerenberg there is much morainic material but, as Wordie reported, trachybasalt of Beerenberg type is frequent in the vicinity of Palffy crater, while three-quarters of a mile to the east massive ankaramite is exposed.

Farther out on the flanks of Beerenberg there are excellent examples of parasitic cones. North-east of the big bend in the Tornoe River (Text-fig. 1) a line of three craters trends approximately N. 30° E. The most southerly vent is about a quarter of a mile from the river and measures 150 yards in diameter. Agglomeratic material occurring on the outer slopes of this crater contains irregular bombs of lava intermediate between ankaramite and Beerenberg trachybasalt, smaller fragments of other types of lava, e.g. trachyandesite, and in certain layers numerous crystals of pyroxene and olivine. The lava extruded from this vent is well exposed in the banks of the Tornoe River where it overlies trachyandesite associated with Scoresby Crater to the south. It is an ankaramite, but it contains a few plagioclase phenocrysts as well as those of olivine and pyroxene. In the northern bank of the river a dyke of ankaramite cuts the lava and is chilled against it. A reduced content of plagioclase serves to distinguish the dyke ankaramite from the lava and permits the recognition of fragments of the latter enclosed in the dyke. The dyke trends N. 18° W. and is almost vertical. If it continues on the same trend beyond the exposures it would pass very close to the most southerly of the three craters and is, therefore, interpreted as a radial dyke associated with this centre, cutting the lavas extruded from the vent. This is the only case on Jan Mayen, as far as the author is aware, where the order of eruption at a single volcanic vent during a single phase of activity can be worked out.

Esk crater, set far out on the lower flanks of Beerenberg, is an unbreached cone with a crater about 300 yards in diameter, which is rimmed by red agglutinate. To the south and south-west ankaramite is exposed which appears to have flowed from Esk crater. A line of old sea cliffs to the south-west of Esk exposes tuffs, agglomerates, and volcanic washes below ankaramite lavas. The cliffs are roughly stratified but the layering is not continuous laterally for any distance and the waterlain washes are of particularly inconstant development. Most of the fragmental material is ankaramitic and it seems likely that these rocks were formed during the first opening of a vent on the site of Esk crater by the settling of pyroclastic material into water. A search for fossils in these washes was unsuccessful. At the south-western end of the line of cliffs the ankaramitic tuff and agglomerate has overwhelmed a pillar 30 feet high of lapilli-tuff which, in its variety of contained lapilli, more closely resembles the lapilli-tuff seen at the base of the cliffs at Cape Fishburn than any other pyroclastic rock at this north-eastern end of the island. The eroded nature of the pillar indicates a considerable lapse of time between the formation of the lapilli-tuff of which it is composed and the deposition of the later agglomeratic material around it.

The agglomeratic rocks are overlain by ankaramitic lavas well

exposed in low cliffs (*ca.* 80 feet high) behind the south-west end of Long Beach where they appear to have breached the pyroclastic deposits. Here, better than anywhere else in north-east Jan Mayen, an extrusive nature for the ankaramite may be demonstrated. Lava falls, tongues of lava draped over irregularities in earlier flows, scoriaceous surfaces and, in one case, tensional tears in the glassy surface skin of a flow remain to testify to the origin of the ankaramites. In every case the glassy skin is roughened by numerous phenocrysts protruding from the lava beneath the skin. There can be no doubt that the ankaramite here is an extrusive rock and that the magma contained the phenocrysts before extrusion.

Behind Turn Bay a recent flow of basanitoid trachybasalt crosses the wide beach from the line of the old sea-cliffs to the present shore. It appears to have originated from a vent to the west of Esk crater and probably represents the most recent phase of activity in the area under study.

Conclusions

The evidence presented in the foregoing paragraphs makes it clear that the volcanic history of N.E. Jan Mayen is more complicated than was previously realized. In this part of the island the following sequence of events can be recognized, all later than the extrusion of trachytes and trachyandesites to the south-west.

1. Formation of lapilli-tuffs by fragmentation of pre-existing rocks not now exposed.

2. Formation of ankaramite agglomerates.

3. Formation of brown tuffs containing bombs of trachybasalt. These first three phases have been recognized in the Fishburn area alone.

4. Formation of ankaramitic agglomerates and tuff-breccias alternating later with extrusion of ankaramite lava, represented by deposits in the Fishburn and East Cross Bay areas.

5. Extrusion of ankaramite lavas probably sub-aerially from the Beerenberg vent, represented by successions at Fishburn, East Cross Bay, around Joris glacier, etc.

6. Extrusion of trachybasalt lavas of Beerenberg type from Beerenberg vent with a few interbedded flows of ankaramite. This was the period of formation of the upper cone of Beerenberg and possibly of minor parasitic activity, e.g. at Muyen's Cape.

7. Pronounced activity on the flanks of the main centre with extrusion of ankaramitic or similar lavas from such craters as Bylandt Rheidt, Esk, etc. There is a tendency for this activity to be related to

fissures sometimes radial to the Beerenberg, e.g. Tornoe River vents, Grönberg ridge of the Fishburn area.

8. Continued sporadic activity on the southern flanks further from the Beerenberg centre with the extrusion of trachybasaltic lavas from relatively recent vents represented by the flows behind Turn Bay and South Lagoon.

The evidence from the East Cross Bay area indicates that the lapse of time between the second and third periods of extrusion of ankaramite (phases 4–5 and 7) was sufficient for the cutting of sea-cliffs several hundreds of feet high in the earlier lavas and agglomerates. This suggests that the two periods are to be referred to different cycles of activity. The interbedding of ankaramite and trachybasalt lavas in the cliff sections of the north-west coast, together with the evidence of a period of erosion between the extrusion of trachybasalts and the later ankaramites at the exposures 3 miles east of Hagerup's Hut, indicates that more time elapsed between the extrusion of the trachybasalts (phase 6) and the third period of ankaramite extrusion (phase 7) than between the second period of ankaramite extrusion (phases 4–5) and the trachybasalt extrusion (phase 6). Phases 4, 5, and 6 are, therefore, grouped as one cycle. This, the second of three cycles, was associated with the Beerenberg centre. The first of the three cycles, represented by phases 1, 2, and 3, is represented entirely by pyroclastic rocks and no evidence is available concerning the centre or centres associated with it. The third cycle (phases 7 and 8) is associated with the parasitic vents on the flanks of Beerenberg.

Each cycle opens with extrusion of ankaramite, either as lava or in a fragmental condition giving rise to pyroclastic deposits of ankaramitic type. The nature of the lava changes during the course of the cycle, the later lavas being trachybasalt. Transitional types between ankaramite and trachybasalt suggest that some close genetic relationship obtains between these various types and it is hoped that future laboratory work will disclose the nature of any such relationship.

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Metamorphism of the Manx Slates

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(PLATE VII)

ABSTRACT

The chief metamorphic minerals are described and, in addition to those reported by Lamplugh, cordierite (at scattered localities along the structural axis) and staurolite (at Glion Darragh, Archallagan) are found to occur. The distribution of metamorphic minerals recorded by Lamplugh is confirmed. The age relation of the new minerals to the movements is deduced from microscopic evidence. This is: early folding and development of biotite; relaxation of stress and development of cordierite; shearing and recrystallization of biotite along planes of fracture-cleavage; high grade metamorphism at Archallagan which outlasted the fracture-cleavage movements.

It is concluded that the postulate of a buried granite beneath the axial belt most readily accounts for the distribution of the metamorphism.

INTRODUCTION

STAUROLITE, garnet, biotite, chlorite, and cordierite are characteristic metamorphic minerals in the Manx Slates.

Almandine, biotite, muscovite, and quartz form the highest grade metamorphic mineral assemblage (amphibolite facies). In addition staurolite has been found in one slide.

Cordierite, biotite, muscovite, chlorite, and quartz are found together and the assemblage belongs to the albite-epidote amphibolite facies.

Biotite, chlorite, muscovite, and quartz are the most widely distributed assemblage and are characteristic of the biotite-chlorite sub-facies.

The lowest grade rocks belong to the muscovite-chlorite sub-facies, and muscovite, chlorite, and quartz are the chief minerals present.

The amphibolite facies is developed only in the south-centre of the island near Archallagan and Foxdale (see Text-fig. 1). Cordierite occurs at scattered localities along the structural axis. It is patchily distributed. The greater part of the island lies within the biotite isograd (biotite-chlorite sub-facies) and only the small remaining areas belong to the muscovite-chlorite sub-facies (Text-fig. 1).

Tourmaline, leucoxene (from ilmenite), and spessartite-almandine garnet, are the chief minerals of metasomatic origin. Tourmaline occurs at scattered localities which in the south of the island are never more than 2¼ miles from the Foxdale granite. It is also concentrated in veins in the Dhoon granite. Leucoxene (as laths) is ubiquitous in a belt along the structural axis (Text-fig. 1). Spessartite-almandine garnet occurs patchily in the Snaefell district. There are local concentrations of apatite.

Description of Chief Minerals

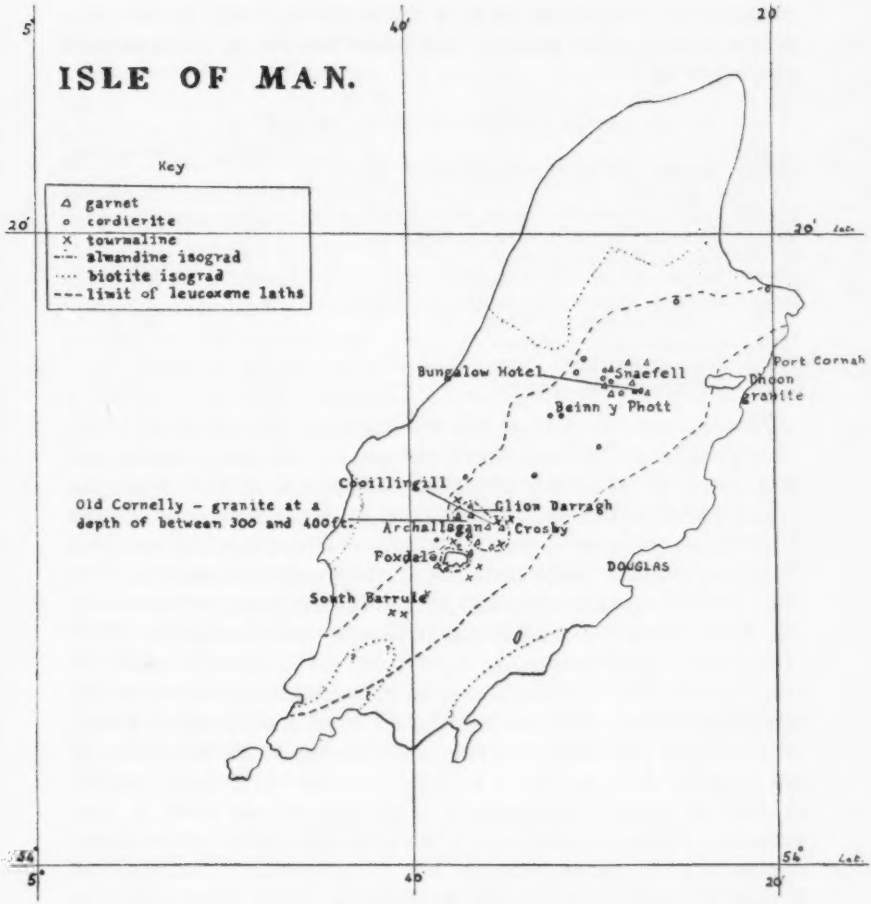
Chlorite.—This mineral is present in slides from all facies. It is frequently an alteration product. It is generally pleochroic from pale green to almost colourless and occasionally shows lamellar twinning. The largest crystals of chlorite occur at Archallagan (0·3 mm.) but generally individual crystals cannot be distinguished. Within the amphibolite facies it occurs as randomly oriented, platy porphyroblasts (0·5 mm.) which are sometimes twinned.

Biotite occurs in all facies except the muscovite-chlorite sub-facies. Within the amphibolite facies it forms plate-like porphyroblasts (0·5 to 1·0 mm.). It is pleochroic from colourless to dark brown (X—colourless, Y—pale brown, Z—dark brown) and contains pleochroic halos. Elsewhere along the structural axis it occurs as a felted mass of small crystals (0·05 to 0·06 mm.) which are pleochroic from light to dark brown. Away from the structural axis the biotite imparts only an aggregate pleochroism (light grey to dark grey, light green to mid green) to the rock. The degree of preferred orientation is high and this imparts flow-cleavage to the rock. The other platy minerals are oriented parallel to the biotite and contribute to the flow-cleavage. The biotite is often partly replaced by chlorite.

Muscovite is best formed at Archallagan where the rock is a silvery schist. Along the remainder of the structural axis recrystallization has been less intense but in places the rock is a silvery phyllite. At Archallagan the muscovite crystals attain a size of 0·5 mm. but elsewhere they are smaller (e.g. less than 0·03 mm.) and occur as a felted mass in the matrix. Where recrystallization has been very slight white mica occurs as a sericitic aggregate.

Almandine garnet occurs as hexagonal porphyroblasts which attain approximately 5 mm. in diameter. It is colourless in thin section but red or pink in hand specimen. There are all stages of replacement by chlorite which is often concentrated marginally. The chlorite is pleochroic from pale green to near colourless. It often contains pleochroic halos. The garnets commonly deflect the foliation forming augen in which interlocking aggregates of quartz are often concentrated. Some of the garnets contain small inclusions which are sometimes ordered and apparently represent former " s " planes. The inclusions are generally in parallel rows which however do not always conform to the existing planar structures in the rock. Inclusions of quartz are common in the garnets and tourmaline is occasionally present.

Spessartite-almandine garnet porphyroblasts attain diameters of 5 to 10 mm. and occur as well-shaped rhombic dodecahedra. They are often partly chloritized but are sometimes quite fresh and unaltered. The chlorite differs from that which replaces the garnets at Archallagan



TEXT-FIG. 1.—Metamorphic Map of the Isle of Man. (Scale: one inch to seven miles.)

in that it does not contain pleochroic halos. The garnets sometimes contain symmetrically arranged inclusions which outline pyramids the apices of which are at the centre of the crystal while the bases of the pyramids are formed by the sides of the crystal. Harker (1952, p. 44) describes this arrangement of inclusions from Mn-garnets elsewhere. The fracture cleavage is deflected by the garnets which pre-date this structure. Leucoxene laths also post-date the garnets and project into pseudomorphed garnet crystals.

In view of the patchy distribution of the garnets in the Snaefell district and in the light of work by Goldschmidt (1921, p. 48), and

others, colorimetric analyses were performed to determine the manganese content of the garnets. The results are set out in the accompanying table.

Garnets from Metamorphosed Manx Slates

		% *MnO*	% *Spessartite.*
1553D.	Garnet, quarry east of Bungalow Hotel, Snaefell	4·38	13·1
1553D.	Dark garnetiferous band, same locality	0·09	
1553D.	Surrounding rock, same locality	0·015	
1519.	Garnet, Snaefell railway	13·39	40·1
1559.	Garnet, north cliffs, Snaefell	5·99	17·9
1609.	Garnet, stream-bed south of Snaefell, north of Beinn-y Phott	16·17	48·4
813.	Garnet, Snaefell railway	12·61	29·2
3790.	Garnet, Foxdale	1·10	3·3
2252.	Garnet, Crosby Quarry	0·92	2·1

These results show a good deal of variation in the amount of MnO. Only 1609 (a specimen from the stream between Snaefell and Beinn-y Phott) would fall within the composition field of spessartite as given by Rankama and Sahama (1949, p. 641). They state that the MnO in spessartite varies from 15 per cent to 40 per cent. All but those from near Foxdale could be regarded as spessartite-almandine. The amount of manganese which the garnets now contain is subject to the alteration which the sample has undergone and the sample which contains the highest percentage MnO is also the least altered specimen.

Specimen 1553D displays a dark band, more garnetiferous than the surrounding slate. Analyses of the dark band and surrounding slate show a higher percentage MnO in the dark band. While the amount is not large in either sample it is likely that the higher MnO-content explains the greater concentration of garnets. Turner (1948, p. 138) states that a high concentration of Mn in the sediment is not necessary to provide a high concentration of Mn in small garnets. Thus although the concentration of Mn in the sediments of 1553D is not large it is likely that there was sufficient to allow the formation of Mn-garnets.

Staurolite occurs as irregular shaped porphyroblasts (0·8 mm.) in Survey Slide E2669. It shows pleochroism from colourless to pale yellow (X—colourless, Y—very pale yellow, Z—pale or mid-yellow). The crystals are sieved with inclusions of quartz. Twinning is displayed by some of the staurolite. Not more than three crystals occur within the section.

Cordierite has a maximum size of 1 to 1·2 mm. within the aureole of the Foxdale granite. Elsewhere the largest cordierite crystals reach 0·6 to 0·8 mm. The mineral occurs both as pseudomorphs and unaltered crystals. Within the aureole of the Foxdale granite the cordierite has been completely pinitized but it often shows its pseudohexagonal crystal form (Plate VII, fig. 1). Elsewhere it is only partly altered or

quite fresh. It sometimes shows twinning and its characteristic pseudo-hexagonal form but often has an irregular or rounded outline.

The cordierite post-dates the flow-cleavage. Traces of this structure remain within the crystals in the form of minute inclusions which lie parallel to the flow cleavage in the surrounding rock (Plate VII, figs. 1 and 2). It is earlier in date than the fracture cleavage which is often deflected by, or continues for short distances within, the boundaries of the cordierites (Plate VII, fig. 2).

In slide 1963 it is interesting that some of the thin veins of quartz which cut the xenoblasts are themselves broken by the fracture cleavage indicating that they post-date the formation of the cordierite and pre-date the fracture cleavage.

Tourmaline occurs as green, lath-like crystals which when the mineral is abundant, can be distinguished in hand specimens with the aid of a lens. It shows pleochroism from light to mid-green (X—near colourless, Z—medium green) in section and sometimes displays a bluish core surrounded by a green border. When most abundant tourmaline is practically the sole constituent of the rock. It displays lath-like crystal form and sometimes occurs in wavering folia. Triangular sections are quite common and the mineral's tendency to idiomorphism where it is of metasomatic origin contrasts with the irregular detrital form which it displays when present as an accessory—e.g. present throughout the Manx Slates.

Leucoxene occurs as plate-like crystals (maximum length about 2 mm.) at the expense of ilmenite which is not always entirely replaced. Its orientation is random and it post-dates the fracture- and flow-cleavage.

Characteristics of Facies

Greenschist Facies.

(*a*) *Muscovite-chlorite sub-facies.*—Muscovite, chlorite, and quartz are the characteristic minerals in rocks which belong to this facies. The grain-size is very small. The quartz grains reach a maximum diameter of about 0·1 mm.; laths of muscovite are about 0·1 mm.; intergrowths of chlorite and muscovite occur which reach 0·2 mm. The chief accessories are green pleochroic tourmaline (0·1 mm.), leucoxene, limonite, black opaque iron, and pyrite. Fracture cleavage is generally present and so also is a weak flow-cleavage.

(*b*) *Biotite-chlorite sub-facies.*—The greater part of the Manx Slates belong to this facies (Text-fig. 1). The recrystallization varies and is greatest along the structural axis. Recrystallization gradually fades towards the outer limits of the facies where it is not apparent in hand specimen. In the outer limits of the zone biotite, chlorite, and muscovite, form a felted mass of tiny crystals which lie parallel to the bedding

or are puckered into microfolds by the fracture cleavage. The biotite aggregate displays pleochroism from near colourless to pale green or from light to dark grey. Along the structural axis the mineralogical composition is generally the same but the crystals are larger. The biotite is pleochroic from light brown to dark brown. Some part of almost every stratigraphic division falls within the biotite-chlorite sub-facies and the proportions of biotite-chlorite and quartz-sericite vary according to the lithological type. Quartz and sericite preponderate in the light bands whereas biotite and chlorite are dominant in the dark bands.

Calcareous nodules occur in the Lonan Flags (stratigraphic division of Manx Slates) in the region of Port Cornah. These are sometimes garnetiferous (Survey slide E2510) and often contain epidote. The mineral assemblage is characteristic of the biotite-chlorite sub-facies to which the surrounding sediments belong but the greater proportion of calcite has allowed epidote and calcite to form, thus distinguishing the assemblage from that of the sediments (Lonan Flags). The apparent selective metamorphism of the nodules is similar to that seen in the Ardennes (Stainier, 1929).

There are patches of garnetiferous slate within the biotite-chlorite sub-facies in the Snaefell district. These occur on Snaefell at the localities described by Lamplugh (1903, pp. 110–111) and in the bed of the stream between Snaefell and Beinn-y Phott where they have been found during the present work. In all but two of the known occurrences the garnets are associated with veins of granular quartz. The veins enclose argillaceous patches which are of the order of 15 cm. across. These contain numerous garnets which average 5 to 10 mm. in diameter, and which are usually chloritized, but sometimes quite fresh and unaltered.

Approximately 900 yards past the Bungalow Hotel up the Snaefell railway a patch of garnetiferous slate occurs. No connection appears to exist between the garnets and near-by quartz veins. The distribution also seems unrelated to either bedding or cleavage. The patch of garnetiferous slate is about 4 feet wide. The concentration of garnets thins out by a rapid reduction in the size and numbers of the crystals and three feet from the edge of the garnetiferous area none can be found either microscopically or in hand specimen. Four slides confirmed these facts. In view of the association elsewhere of the garnets with quartz veins it is particularly interesting that a narrow vein of chlorite, quartz, and muscovite, not more than three feet from this garnetiferous slate, is quite barren of garnets as is the surrounding slate.

In the small quarry, previously described by Lamplugh (1903, p. 110), to the east of the Bungalow Hotel, garnets are to be found. Their

distribution is once more distinctive and independent of quartz veins though influenced by the bedding since they are more abundant in 1 cm. thick bands of dark grey slate than in the remainder of the rock. Specimen 1553D (see p. 144) was obtained in this quarry.

Albite-epidote Amphibolite Facies.

Within the biotite isograd patches of cordierite-bearing rocks occur. The mineral assemblage in these rocks, muscovite, chlorite, biotite, quartz, cordierite, is characteristic of the albite-epidote amphibolite facies.

Amphibolite Facies.

(*a*) *Cordierite-anthophyllite sub-facies.*—The development of this facies is strictly localized and only occurs within a few hundred feet of the Foxdale granite. It is illustrated by the following slides :—

Survey Slide E2889.—This slide shows a contact between the granite and surrounding hornfels. At the contact there are porphyroblasts (1·5 mm.) of colourless garnet but those farther from the contact (though in the same section) are smaller (0·4 mm.) and less well formed. Apatite enrichment has taken place within about 1 mm. of the contact where this mineral comprises about 50 per cent of the rock. The apatite occurs as uniaxial negative, anhedral crystals (0·08 mm.). Accessory amounts of apatite occur within the granite and within the hornfels situated farther from the contact. Tourmaline is enriched within a zone about 1 mm. thick from the contact with granite. It comprises about 10 to 15 per cent of the rock within this zone but elsewhere, both in granite and hornfels, it is only an accessory constituent. There are platy porphyroblasts of biotite (0·75 mm.) which show pleochroism (X—near colourless, Y—deep brown, Z—reddish brown) and often contain pleochroic halos. They are sometimes partly replaced by pale green pleochroic chlorite. The biotite shows little preferred orientation. The matrix is composed chiefly of small (0·1 mm.) interlocking quartz crystals.

1892, 200 feet west of granite contact.—Small garnets (0·8 mm.) are present which are largely pseudomorphed by pale green, pleochroic chlorite. Cordierite occurs as hexagonal porphyroblasts (1 mm.) now pseudomorphed by faint yellow-green pinite. There are randomly oriented crystals of pale green, pleochroic chlorite (0·3 mm.) which sometimes show lamellar twinning. They often contain pleochroic halos, in which respect they resemble the partly replaced biotite porphyroblasts of E2889. It appears possible that these crystals of chlorite were originally biotite porphyroblasts now completely replaced. The matrix consists of folia of chlorite and sericite and interlocking quartz crystals (0·12 mm.).

Survey Slide E2803.—This slide contains similar pseudomorphed cordierite porphyroblasts (0·75 mm.) to those seen in slide 1892.

Turner (1948, p. 79) explains the presence of both almandine and cordierite in this facies by reference to the limited extent to which Mg and Fe may replace one another in these two minerals so that the two elements behave as independent components.

(*b*) *Staurolite-kyanite sub-facies.*—There are few exposures for about ¾ mile to the north of the Foxdale granite but one small exposure indicates that the rock is not garnetiferous (slide 1887). ¾ mile to the north-north-east, exposures of garnetiferous mica schist occur at Archallagan where the metamorphism in the Isle of Man reaches its highest grade. The metamorphism increases rapidly from all sides towards this locality and this increase is most clearly seen in the beds of the streams (Cooillingill and Glion Darragh) which drain northwards from Archallagan. Exposures to the east, west, and south, of Archallagan are not sufficiently plentiful to allow the garnet isograd to be positioned more accurately than to within about ¼ mile. To the north, in Glion Darragh, exposures are nearly continuous and the isograd may be defined to within 50 to 100 feet.

Survey slide E2669, western branch of Glion Darragh (Lamplugh, 1903, p. 112), contains staurolite. The occurence is evidently very localized since despite the preparation of several more slides from specimens obtained in the same area no further occurrence of staurolite has been found.

The general microscopic appearance of these schists is of alternating folia of interlocking quartz crystals and parallel oriented platy minerals. The schistosity is deflected by hexagonal porphyroblasts of garnet which form augen. Quartz is often concentrated in the corners of the augen formed by the garnets. Plate-like crystals of brown, pleochroic biotite (0·75 mm.), with pleochroic halos, are of irregular distribution. Tourmaline, leucoxene, magnetite, sphene, and zircon form accessories.

Relation of the Metamorphism to Movements

By reference to the most important cleavages in the rocks it is possible to distinguish definite phases in which certain characteristic new minerals appeared and hence to correlate stages in the metamorphism with the movement phases.

Muscovite-chlorite and Biotite-chlorite sub-facies.—The earliest movements led to sharp folding and the development of steeply inclined flow-cleavage. This is imparted to the rock by platy minerals such as chlorite, muscovite, and biotite which formed during these movements by recrystallization. During this movement phase it is unlikely that the grade of metamorphism was anywhere advanced beyond the biotite-chlorite sub-facies and in regions away from the

structural axis towards both the south-east and north-west sides of the island very little mineralogical change took place. This stage of metamorphism is the most widespread and typical of the Manx Slates but in places it has been obscured by later metamorphism and metasomatism.

Spessartite-almandine garnets.—The fracture-cleavage is deflected by these garnets and evidently they developed before this latest phase of movements. Their relationship to the flow-cleavage is less apparent. This cleavage generally stops short at the margins of the crystals and does not appear to flow around them (except where dragged into that relationship by the fracture cleavage), as would be the case if the cleavage post-dates the crystals. Therefore it is likely that the garnets post-date the early movements.

Albite-epidote amphibolite sub-facies.—Cordierite occurs at scattered points along the structural axis and appears to have formed later than the biotite of the flow-cleavage movements since trails of inclusions lying parallel to the schistosity can frequently be traced across individual crystals. The planes of fracture-cleavage either penetrate a short distance into the cordierite crystals or are deflected by them. The cordierite therefore formed prior to the movements in which the fracture-cleavage developed.

Cordierite is listed by Harker (1952, p. 231) as a typical " antistress " mineral by virtue of its common development under conditions of low-grade purely thermal metamorphism, but in Scotland it is also known to occur under " regional " conditions. Harker attributes its presence to deficient shearing-stress and emphasizes that its occurrence is quite distinct from a second metamorphism. The " widespread and abundant " distribution of this cordierite contrasts with that in the Manx Slates, which is patchy. It seems unlikely that maximum shearing-stress could prevail throughout the greater part of a series but have a reduced value at scattered localities. If the temperature of the rocks remained elevated after the reduction of shearing-stress one would anticipate a general development of cordierite provided the rocks were, in general, of appropriate chemical composition. The Manx cordierite is patchily distributed but there is no evidence to suggest that this is dependent on changes in lithology such as might have caused the chemical composition to be patchy. The scattered development of cordierite may have been caused by the local introduction of appropriate elements required for its development (e.g. MgO possibly). There is no evidence to support the view that shearing-stress was below its maximum value during the early movements and early metamorphism and therefore an introduction of, say, Mg during the early movements would have been unlikely to lead to the development of cordierite. Hence it seems probable that the cordierite developed in response to the metasomatic introduction of certain elements at a period when the

early movements had finished or become reduced in intensity. It seems likely that the cordierite developed at a period of reduced stress between the major movements.

Amphibolite facies.—Fracture-cleavage is present outside this facies but gradually becomes indistinct as the garnet isograd is approached. It seems probable that recrystallization outlasted the fracture-cleavage movements.

Hydrothermal Action

Cause for Patchy Distribution of Spessartite-Almandine Garnet.—In many areas garnet forms the index mineral for the next higher zone to that of biotite in pelitic and semi-pelitic rocks subjected to regional metamorphism. It was found by Goldschmidt (1921, p. 48) that in the Stavanger district of Norway the order of appearance of biotite and garnet was reversed and garnet developed before biotite. This garnet was shown by him to be spessartite-almandine containing 12·28 per cent MnO.

Many authors have since described the development of Mn-garnets under conditions of low-grade metamorphism. It has been shown (Turner, 1948, p. 138) that a high concentration of Mn is not required to allow a considerable development of garnet. Garnetiferous quartz veins, evidently similar to those in the Snaefell district, have been described by Emmons (1918, p. 52), and by Barth and Balk (1936, pp. 684–774).

It seems probable that Mn was introduced into the Manx Slates of the Snaefell district by certain of the quartz veins and this allowed spessartite-almandine garnet to develop. Possibly the Mn was introduced by veins of only a certain age (quartz veins both pre-date and post-date the fracture-cleavage) and this perhaps accounts for the absence of garnets from the majority of quartz veins.

Boron Metasomatism.—Tourmaline is concentrated in veins in the Dhoon granite and at localities in the south of the island which are never more than 2¼ miles from the Foxdale granite.

Lamplugh denied any relationship between the Foxdale granite and the tourmaline on the grounds that the tourmaline was too far from the intrusive for it to have been responsible (1903, p. 113, 169). But Hutton (1939, pp. 559–602) attributes regionally distributed tourmaline to boron-carrying magmatic solutions even when the tourmaline occurs as much as 120 miles from any known outcrop of granite. In the more highly metamorphosed equivalents of these rocks tourmaline becomes increasingly abundant.

In the Manx Slates, when tourmaline is a major constituent of the rock, it seems justifiable to conclude that it is of metasomatic origin and is probably related to granite. The accessory tourmaline is likely

to have been an original constituent of the sediments as the crystals often display detrital form.

Causes of the Metamorphism of the Manx Slates

(*a*) *Regional Metamorphism.* (i) *Lamplugh's Views.*—Lamplugh attributes the metamorphism to " differential movements " but admits that the granites may " have played their part in raising the temperature of the rock-mass and bringing it nearer to the critical stage " (1903, p. 114). The development of new minerals (e.g. garnet on Snaefell and tourmaline south of South Barrule) in the vicinity of quartz veins is attributed to " increased crushing and friction developed along the junction of soft and hard masses " (1903, p. 109). Recrystallization is stated to be " chiefly prevalent in the banded strata of transitional type which lie between the Barrule Slates and Grits " (1903, p. 108).

He suggests that there is a relation between the distribution of the breccias and metamorphism with respect to the structural axis, and states that " it is noteworthy with the greatest degree of alteration is attained where the planes of dominant ' shear cleavage ' are flat or nearly so " (1903, p. 114).

(ii) *Discussion of Lamplugh's Views.*—The association of new minerals with quartz veins indicates that these minerals are of hydrothermal origin.

The present work indicates no greater degree of recrystallization in banded strata than in homogeneous slate.

The origin of the breccias is unlikely to have any relationship to the metamorphism and in any case there is no real comparison between the distribution of the breccias and metamorphism.

The most widespread metamorphism pre-dates the fracture-cleavage which has a shallow dip along the structural axis. The more intense recrystallization at Foxdale and Archallagan appears to have outlasted the fracture-cleavage movements. The fracture-cleavage probably has no direct bearing upon the causes of the metamorphism.

(iii) *Alternative Hypothesis.*—The coincidence of the first phase of metamorphism with the early movements at first sight seems to support the view that the movements caused the metamorphism but as metamorphism dies out towards the north-west and south-east, where these movements were no less intense than in the metamorphic region, the conclusion is not supported.

The Lonan Flags exposed to the south-east of the metamorphic belt are the oldest rocks of the Manx Slates Series whereas to the north-west of the metamorphic belt the youngest beds of the series are exposed (Gillot, 1955). Therefore it is not possible to explain the difference in metamorphic grade, as between the south-east and north-west coasts

and the axial belt, as due to an insulating effect dependent on thicker sedimentary cover.

While the conclusion that deformation exerts a catalytic effect on metamorphic reaction is not questioned, in the Manx Slates, some other agency in addition seems required to explain the development and distribution of new minerals.

Stainier (1929) has postulated that a granitic batholith underlies the axial region and that the Foxdale and Dhoon granites are apophyses of the deeper granite. If the postulated granite, beneath the Manx Slates, is assumed to approach most closely to the surface below the axial belt the difference in the degree of metamorphism between that belt and the north-west and south-east districts of the island may be explained. In my opinion this view meets with the fewest difficulties as an explanation for the distribution of Manx metamorphism.

(*b*) *Localized high-grade metamorphism.* (i) *Lamplugh's Views.*—Granite of Foxdale type was found in the shaft of the Old Cornelly lead mine, Archallagan, at a depth of 300 to 400 feet (Lamplugh, 1903, p. 111). Lamplugh does not attribute the metamorphism in the Archallagan region to the heating effect of the granite, on the grounds that comparable metamorphism is lacking in the sediments which surround the Foxdale granite itself. The metamorphism at Foxdale differs in kind, but not degree, from that at Archallagan.

(ii) *Alternative to Lamplugh's View.*—Brindley (1948, pp. 27–28) reports that hornfels occurs near the contact of the Leinster granite and as isolated patches within the massif, but away from the contact the hornfels passes into mica-schist, which constitutes the main rock-type of the aureole. This again merges into slate farther out. Brindley considers that the schist and hornfels are strictly contemporaneous, the schist having developed under intense paracrystalline deformation within the granitic aureole, the whole of which represented a mobile or plastic envelope to the igneous body. Contact metamorphism took place hand in hand with the intrusive movements.

This explanation is not exactly applicable to the metamorphic high in the Foxdale-Archallagan district of the Isle of Man since no gradual outward passage from hornfels into schist occurs in the rock surrounding the granite but it suggests an explanation which may fit the facts of the case.

If the rocks exposed at Archallagan were subjected to shearing movements while the rock surrounding the Foxdale granite was in some way protected, the development of schist at Archallagan and hornfels at Foxdale would be a natural consequence. No information is at present available as to the direction from which the granite was intruded or as to whether the Foxdale and Old Cornelly granites are directly connected but it seems possible that the Cornelly granite

underwent either more intense, or more protracted movements than the Foxdale granite and hence exerted a greater shearing effect upon the overlying sediments. It may be that the difference is connected with the different relative positions of the sediments at the two places; at Foxdale they lie at the same structural level as the granite while the roof has been removed by erosion, but at Old Cornelly the schist lies 300 to 400 feet above the intrusion towards which its greatest intrusive force may have been directed.

In any event and in view of the inadequacy of other explanations, it seems to me most reasonable to attribute the local high-grade metamorphism in this region to the heating effect of the granite.

ACKNOWLEDGMENTS

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EXPLANATION OF PLATE VII

FIG. 1.—Slide 1892 * (× 38), quarry 200 feet W. of Foxdale granite contact. Pseudomorphed cordierite porphyroblasts. Flow-cleavage is traceable within the cordierite which has partly assimilated the groundmass, and the cordierite post-dates the flow-cleavage.

FIG. 2.—Slide 1963 * (× 21), East Baldwin stream. Xenoblastic cordierite porphyroblasts in a rock along the structural axis of the Manx Slates. The porphyroblasts post-date the flow-cleavage, but the fracture cleavage has been deflected by the cordierite or only penetrates a short distance into the crystal.

* Numbers refer to slides in the Geology Dept., University of Liverpool.

FIG. 1.

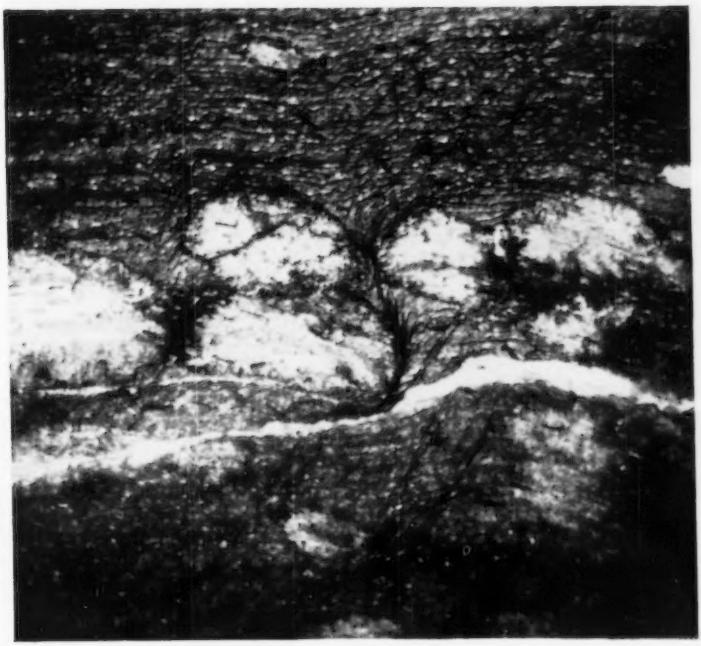


FIG. 2.

Flow Phenomena in a Limestone on Changalumi Hill, Southern Nyasaland [1]

By M. S. Garson

Abstract

A Basement Complex graphite-limestone on Changalumi Hill, along the eastern escarpment of the Southern Nyasaland rift-valley, exhibits various phenomena including contorted exotic boulders, migration of graphite, and injection of discordant veins into adjacent gneiss. These features indicate that the limestone has burst through its anticlinal roof during strong isoclinal folding at a relatively low temperature, engulfing and distorting a minor synclinal infold. The age of the break-through is proved to be pre-Chilwa Series and pre-Rifting.

Introduction

THE Changalumi limestone was first briefly described by Dixey (1927, 29), who, however, made no reference to any unusual features. Later Cooper (1955) described the limestone as a flow-limestone, possibly due to migration from a deep metamorphic zone. The writer recently mapped the area in connection with a survey for limestones suitable for cement production. The many anomalous features in the limestone provided valuable clues as to the structure of the hill, and the economic question of increase or decrease in boulder-content in depth.

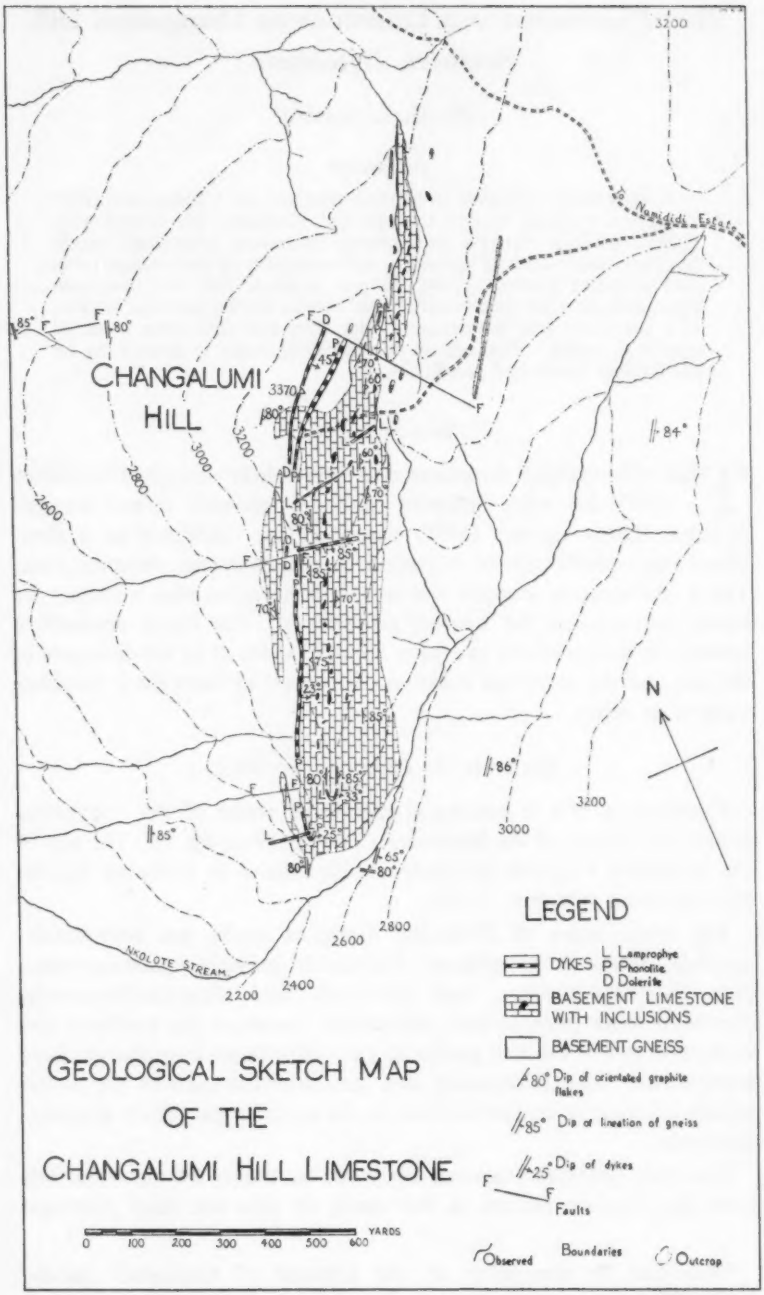
Geology of Changalumi Hill

Changalumi Hill is composed mainly of rocks of the " graphite-limestone " facies of the Basement Complex (Text-fig. 1). The age of the Basement Complex in Southern Nyasaland is probably Middle Pre-Cambrian (Cooper, 1955).

The main types of Basement Complex rocks are hornblende-gneisses, biotite-garnet-gneisses, feldspathic gneisses, quartz-gneisses, diopside-sphene-gneisses, and graphite- and diopside-limestones. The hornblende-gneisses and migmatites, found on the northern and western slopes of the hill, gradually pass into diopside-sphene-gneisses southwards. The feldspathic and garnet-biotite-gneisses are found mainly adjacent to, and as boulders in, the northern part of the graphite-limestone.

The main graphite-limestone body can be traced for almost a mile from the Nkolote Stream in the south, to near the deep east-west

[1] Published by permission of the Director of Geological Survey, Nyasaland.



TEXT-FIG. 1.—Geological sketch map of the Changalumi Hill Limestone.

gully at the north of the hill. Several minor parallel limestone bands occur close to this large outcrop.

The most conspicuous feature of the large graphite-limestone belt is the occurrence of boulders of all types of country gneiss, embedded in the limestone, chiefly along the crest-line. These boulders, which measure from inches across up to over 100 feet long and 50 feet across, so crowd the limestone at its northern end as to merit giving some of the outcrop the name of limestone-breccia. Towards the southern end of the limestone the size of these boulders diminishes, the limestone becoming more coarsely crystalline, and containing less graphite and more green diopside.

The general strike of the Basement Complex rocks is along a N.N.E.–S.S.W. line. Dips of foliation are invariably steep and towards the E.S.E., with the exception of anomalous dips in the limestone.

The hill has been intruded by later minor alkaline and basic dykes of the Chilwa Series (Dixey, Campbell Smith, and Bisset, 1937). The age of this series is considered by Dr. Dixey to be immediately post-Karroo, i.e. Rhaetic to Lower Lias.

These intrusives include a fine-grained phonolite dyke and a 3-foot dolerite, parallel to the western boundary of the limestone and dipping eastwards into the hill, and several thin vertical lamprophyre dykes of camptonitic type, aligned in an east-west direction. These dykes cut cleanly through the limestone and exotic boulders, indicating that the emplacement of the latter occurred in pre-Chilwa Series times.

The principal faulting in the area is of rift-valley type along a N.E.–S.W. line; a line of weakness emphasized by the entrenchment of the V-shaped Nkolote Stream valley. This major fault-line brings the limestone up against shattered basic gneisses in the Nkolote Stream. Small east-west faults on Changalumi Hill, which slightly offset the limestone and dykes, probably also belong to the rifting episode.

Petrology of the Changalumi Limestone

The Changalumi Hill graphite-limestone is essentially similar, apart from its boulder-content, to many other graphite-limestones in Southern Nyasaland. The limestone is white to grey and pinkish in colour, weathering to a dull bluish-grey. The texture is coarsely granular to saccharoidal in the northern part of the outcrop, becoming coarser towards the south, where a strong development of larger euhedral calcite crystals gives the rock a pegmatitic appearance. The calcite grains are strongly twinned in nearly all cases, except in the reaction zones at the dyke margins. The twinned crystals are often bent, and show cataclastic strain shadows.

Orientated graphite flakes, up to a few millimetres long, which run

to a few per cent in much of the northern part of the outcrop, in places give the limestone a gneissose appearance.

On the weathering surface small knots of greyish to pinkish material are prominent over much of the outcrop. In thin section these are seen to be partly wollastonite, and partly feldspar from the disintegration of feldspathic gneiss boulders. Small rounded quartz grains are also present, probably originating from the breaking up of quartz-gneisses, as any original silica is likely to have been taken up into calc-silicate minerals.

The magnesian-rich minerals are found in the flanks of the southern part of the outcrop as greenish bands, in which diopside may amount to 10 per cent. The carbonate mineral throughout the outcrop is dominantly calcite, so that the magnesian content is almost entirely concentrated in silicate minerals.

Other minerals identified in thin section are reddish-brown sphene up to 1 cm. long, rare muscovite and magnetite, and pyrite oxidized partly to limonite.

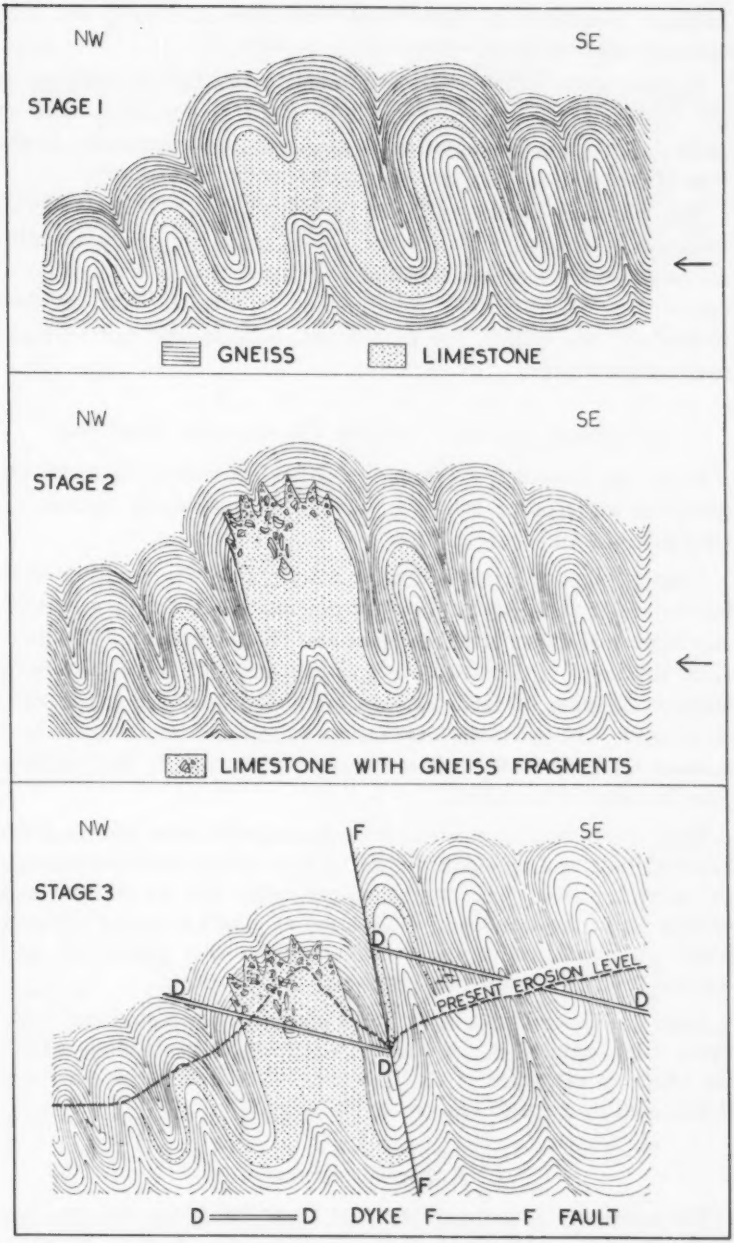
Flow Phenomena and Description of Contorted Exotic Boulders

Evidence of flow movement is perfectly illustrated at the south-western extremity of the Changalumi limestone, where several thin limestone veins connected with the main body, intricately cross-cut and penetrate the adjacent diopside-sphene-gneisses. These veins dip abruptly W.N.W. through the normal steep E.S.E. lineation of the gneisses.

Lines of orientated graphite flakes, and trails of fragments from shattered gneisses, also show anomalous W.N.W. dips along the western half of the limestone.

Striking illustrations of flow in the limestone are accorded by the included boulders. The larger boulders, which are up to 100 feet long and 50 feet across, are generally elongated along the strike of the limestone, but in some cases show no symmetrical orientation to the latter. Where these boulders consist of banded material it can be seen that the dip of lineation is seldom that of the adjacent country gneisses, but usually inclined haphazardly in all directions. The edges of these larger boulders are rounded, probably due to attrition and rolling during their movement in the limestone.

The small boulders and slabs of gneiss offer the most spectacular evidence of flow. Boulders up to a few feet across are often split open and pierced by veins of limestone, or are stretched out and broken into trails of debris. Thinner slabs of gneiss, which have been peeled off adjacent larger masses, are indescribably folded and contorted by differential flow-movement in the limestone. Several examples were seen of thin bands of gneiss, which have been folded



TEXT-FIG. 2.—Schematic diagram of the Structural Evolution of the Changalumi Limestone.

back on themselves to form cylindrical open structures, and then subsequently " rolled ".

In many cases a thin concentration of graphite has accumulated on the " windward " or concave bottom edges of contorted slabs of gneiss. This is probably due to migrant graphite being caught by the trap-like structures of these sluggishly-moving boulders.

The presence of the relatively low-temperature index minerals wollastonite and diopside (Bowen, 1940), probably connected with the original regional metamorphism anyway, and strongly-twinned calcite showing cataclasis, would seem to indicate that the flow phenomena are mainly due to dynamic reasons, and not to high-temperature causes.

Geological History of the Changalumi Limestone

From the foregoing evidence the writer accordingly suggests the following sequence of events to explain the anomalous features on Changalumi Hill (Text-fig. 2).

Stage 1.—Strong pressure from an easterly direction produced large isoclinal folds, dipping steeply to the east and south-east, in an originally horizontal sequence of limestone and other sediments. The limestone thickened in a bulbous fashion at the top of an anticline with attenuated limbs. A small isoclinal infold was produced in this anticlinal crest. The sediments were regionally metamorphosed to various gneisses ; and diopside, wollastonite, etc., crystallized in the limestone from its original impurities.

Stage 2.—Increasing pressure from the same direction tightened the isoclinal folds, and caused the limestone to penetrate and burst through the overlying rock. The small synclinal infold was dragged out and broken up in the intrusion. Thin slabs of gneiss from the adjacent rocks and remnants of the synclinal infold were peeled off, and " rolled ", in the upward movement of the limestone.

Stage 3.—During the Chilwa Series episode thin alkaline and basic dykes were intruded into the area. Later rift-valley faulting cut off the southern extension of the limestone. The present erosion-level is indicated by a dotted line on the diagram.

Acknowledgment

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The Stratigraphical Range of the Foraminiferal Genus *Orbulina* D'Orbigny 1839

By R. N. C. Bowen

Abstract

This genus is not present in pre-Oligocene rocks and probably does not occur below the basal Miocene. Stratigraphical and micropalaeontological evidence is produced to support this statement.

Introduction

THE foraminiferal genus *Orbulina* was founded by d'Orbigny in 1839, the type species *O. universa* being described from recent sands and recorded from numerous, widely separated localities such as the Adriatic coast of Italy near Rimini and the shores of Cuba. The generic diagnosis noted a free, spherical, hollow test perforated by many minute pores and a small round aperture. Subsequent investigations by Pourtales and others, established that the globular end chamber characteristic of the genus represents a later ontogenetic stage, the life history commencing with a small, trochoid *Globigerina* which becomes partly or wholly enclosed and occasionally resorbed. Equipped with fine, elongate spines (not preserved in fossil specimens), the genus is well adapted to a planktonic mode of life. Thick-walled specimens are believed to adopt a benthonic habit during life and with this is associated the development of two or more distinct layers in the walls. It may well be, however, that the cytoplasm in each individual was sufficiently abundant to render it buoyant, since some of these specimens have been found associated with tropical planktonic assemblages of other foraminifera. One species, however, namely *O. imperfecta* Rhumbler has only been found in bottom samples, and must be regarded as definitely benthonic. Brady (1884) described in detail the morphological features of recent specimens of *O. universa*, but was unable to find any apertures except those caused by damage. Coryell and Mossman (1942), while noting the absence of definable apertures in abundant specimens of this species from the Pliocene of Panama, remarked that " it might well be that this genus should be divided on the presence or absence of this simple apertural feature ". Glaessner (1945) stated that there is " no aperture in the last chamber " and Cushman (1948) was " inclined to agree with Earland that the so-called larger aperture of *Orbulina* is an accidental opening ". In 1951, Bronnimann observed that there is " no definite aperture present ", and emended the definition of the genus to include *Candorbulina* Jedlitschka 1934. This form (type species *C. universa* from the Miocene of Czechoslovakia) with circular pores along the sutures

separating the early *Globigerina* portion from the almost spherical end chamber, grades into the distinctly spherical test without projecting *Globigerina* chambers of the true *Orbulina*. It is not always possible to discern these openings due to infilling or masking by adherent matrix. Bronnimann included multichambered individuals from the Czechoslovak Miocene, assigned to the genus *Candeina* d'Orbigny, 1839, by Jedlitschka, 1934, in the genus *Orbulina*, since they differ structurally and in texture from the typical species of *Candeina*. *O. universa* was first recorded fossil from the Tertiary of the Vienna Basin by d'Orbigny in 1846. Subsequently, a very large number of fossil records were made by other authors and in addition, numerous other species have been founded, some of Tertiary age, some earlier. Among the latter are *O. intermedia* Matthew, 1895, a form having a pitted surface often with ridges or rugosities, a diameter varying from 1 to 1·25 mm., obtained from the Lower Cambrian of New Brunswick, Canada ; *O. permiana* de Gregorio 1930 obtained from the Permian of Sicily ; *O. vesca* Zalessky 1926, a spherical form with a diameter of from 0·028 to 0·1 mm., obtained from the Jurassic of the Mine de Kachpoursk, Simbirsk, U.S.S.R., etc. From descriptions and figures of these and other pre-Tertiary species introduced by Haeusler, Karrer, Kübler and Zwingli, Terquem and other authors, it is evident that they lack the necessary generic characters, and do not belong to the genus *Orbulina*. It is now generally agreed that all pre-Eocene specimens are globoid objects, incorrectly assigned to this genus, and Glaessner (1945, p. 149) asserted that " most . . . records of pre-Miocene occurrences . . . are erroneous ". This paper resulted from an investigation of his statement by the present author.

Analysis of Material

Specimens were examined from British Upper Cretaceous and British and American Eocene localities. Williams-Mitchell (1948) recorded *O. universa* from " the whole range of the Chalk in varying quantities ", and regarded it as a useful zonal guide " particularly in the Middle Chalk and in the higher beds of the Upper Chalk ". Specimens from the Red Chalk have been examined and found to be closely similar to those mentioned by Williams-Mitchell. However, they cannot be assigned to the genus *Orbulina* since they lack the characteristic structure and are usually much too small. There can be little doubt that they are inorganic in nature.

During an examination of the foraminiferal fauna of the London Clay carried out from 1950 to 1952, the author identified four specimens as *O. universa*. In view of the small number of specimens available, visual inspection only was possible, but further specimens have been

found subsequently and more detailed examination made in consequence. After sectioning, observation in transmitted light, etc., it has been found that the identification was in error, and the specimens incorrectly assigned to the genus *Orbulina*.

In 1943, Beck recorded *O. universa* from the massive grey mudstones of Eocene age which outcrop along the west bank of Cowlitz River (one and a half miles east of Vader, Lewis County, Washington, U.S.A.) and reach a thickness of about 200 feet.

He referred to the species as " common ", but added that " it is not represented in any sample by a great number of specimens ". He could not observe any aperture in the specimens, and concluded that it was " obscured ".

In 1954, Beck sent the author six paratypes identified by him as *O. universa*. Five of these were complete and one broken. The latter lacks any internal structure, and is very much thicker-walled than is usually the case in this species. Four of the other five were sectioned and found to be thin-walled, but without the inner *Globigerina* chambers characteristic of the genus. All the specimens had diameters in the region of 0·3 mm. The author is of the opinion that none can be assigned to the genus *Orbulina*.

Through the co-operation of Mr. A. G. Davis, it was possible to select all Palaeogene specimens identified as *Orbulina* from the collections of the British Museum (Natural History). Twenty-nine were examined, most of them being of Eocene age. Eleven were obtained from London Clay collections, two from Bracklesham beds collection, one from a collection of material from the French Calcaire Grossier (Lutetian), nine from the A. S. Piret collection (1909) of material from the Wemmelian (Bartonian) of Brabant, Belgium, and six from Tertiary deposits near Cracow, Poland. The largest specimen was one of those obtained from the Bracklesham beds. It was 0·875 mm. in diameter, but after sectioning had been carried out, it was found to be a solid sphere of calcite. The remaining specimens were smaller, ranging from 0·1 to 0·6 mm. in diameter and from spherical to ovoidal in shape. Most of them were sectioned and found to be either solid or thick-walled and devoid of internal *Globigerina* chambers. The specimen from the London Clay figured by Sherborn and Chapman (1886) could not be sectioned, but its very small size (diameter of 0·14 mm.) and the external appearance indicate that it cannot be regarded as a true *Orbulina*. It is registered as P 3722. The rest of the London Clay specimens are of similar size except for P 12323 and P 9810, which are much larger, being 0·36 and 0·45 mm. in diameter respectively. The Belgian specimens are moderately large and mostly ovoidal in shape. The Polish specimens are all spherical, and usually about 0·4 mm. in diameter. The other specimen from the Bracklesham

beds (both specimens are from Box XXX of the Rupert Jones collection in the British Museum (Natural History) and lack registration numbers) is 0·6 mm. in diameter, and that from the Calcaire Grossier is 0·33 mm. in diameter (and again unregistered). None of these specimens possesses the diagnostic characters of the genus *Orbulina*, and all have been misidentified. As a result of this investigation, the author is convinced that the genus does not occur in pre-Oligocene rocks.

Pre-Miocene Records of *Orbulina*

Examination of Cretaceous and Eocene specimens by the author and others has shown that none can be regarded as true *Orbulina*. Numerous records of the genus have been made from Oligocene deposits, however, and the validity of these must be called in question in view of recent work. Eames (1953) has demonstrated that the ages of beds in certain parts of the world need reconsideration, and that confusion has arisen in regard to the use of the term " Aquitanian ", some authors applying it to Lower Miocene deposits, others (including many continental and American workers) to beds of Upper Oligocene age. Oligocene records of the genus *Orbulina* are too numerous for a complete list to be attempted here.

Based on information received from other micropalaeontologists, the stratigraphical figure in LeRoy (1952) shows the lowest limits of *O. universa* as Middle Oligocene in North-Central Venezuela and Upper Oligocene in Ecuador and Trinidad, B.W.I.

Elsewhere, they are given as Lower Miocene (Southern Turkey and Algeria), Middle Miocene (Central Sumatra as determined by LeRoy himself) and the top of the Upper Miocene (Pacific coast).

The existence of an *Orbulina*-surface in Middle Tertiary sections throughout tropical and sub-tropical zones of the world was suggested by LeRoy (1948). It represents a time-datum formed by points of lowest stratigraphic occurrence of *O. universa* in continuously deposited, deep-water globigerine facies deposits. It was criticized by Bronnimann (1951), who stated that it would be invalid if " it should be established that the first appearance of *Orbulina* occurred at different times in different localities within the same region ".

The Oligocene records of *Orbulina* from the Caribbean area are suspect in view of recent work. Eames (1954) pointed out that " the upper part of the so-called ' Oligocene ' in the Caribbean region is probably of Lower Miocene age ". The important palaeontological change which occurred between the Oligocene and the Miocene actually takes place well below the " Oligocene-Miocene " boundary as interpreted until recent years. It seems likely that the age of the *Globorotalia fohsi* biostratigraphic unit from which Bronnimann recorded *O. suturalis* Bronnimann, *O. bilobata* (d'Orbigny), and

O. universa d'Orbigny may be Lower Miocene as Kugler (1953) stated, and not Upper Oligocene as he gave it.

Conclusions

It appears probable that pre-Miocene records of the genus *Orbulina* fall into two distinct categories.

First, a group based upon misidentifications. Examples of this group have already been cited. A significant point is that misidentified objects are usually present in very small numbers. These are incompatible with the breeding capabilities of a pelagic genus such as *Orbulina* which is often found in "floods" in post-Palaeogene deposits. Second, a group recorded from the Caribbean Oligocene. The confusion with regard to the Oligocene-Miocene boundary in this area has been examined above, and the Lower Miocene age of the Caribbean "Upper Oligocene" is being recognized by a growing number of micropalaeontologists.

The *Orbulina*-surface is thus shown to be confined to the Middle Tertiary, though possibly not strictly contemporaneous in its lateral extent, and the genus *Orbulina* is extremely valuable as a Neogene zone-fossil.

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The author is indebted to Dr. F. E. Eames for access to the foraminiferal collections of the Anglo-Iranian Oil Company, Beaufort House, London, and for stimulating discussions on the subject of this paper. Thanks are also due to Mr. A. G. Davis for permission to select and examine specimens from the foraminiferal collections of the British Museum (Natural History).

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Additions to the British Eocene Brachiopod Fauna

By G. F. Elliott

Abstract

Two species of *Argyrotheca* new to this country are recorded from the type Bracklesham Beds, and a full description given of the associated *Terebratulina edwardsi*, previously known from immature specimens only. The British Eocene brachiopods are listed, and the ecological significance of their rarity and distribution briefly discussed.

Introduction

RECENTLY four new species of brachiopods were described from the Eocene of this country, including the first British records of *Argyrotheca* at this horizon (Elliott, 1954). After publication, Mr. D. Curry kindly placed at my disposal a tube of brachiopod material picked from his washings of the Bracklesham Beds (Bed Fisher 21,[1] Auversian), Selsey, Sussex. In this material remains of at least two further species of *Argyrotheca* occur; also specimens permitting a description of the adult shell of *Terebratulina edwardsi* Elliott, described from an immature specimen after having been represented thus in collections for many years. The material has been deposited in the British Museum (Natural History), Dept. Geology, registered nos. BB.13093–13104.

New Records

Terebratulina edwardsi Elliott

Elliott, 1954, p. 722, pl. 15, fig. 3

Diagnosis.

Shell about 7·5 mm. long and 5·5 mm. wide; biconvex, rounded-pentagonal in outline, finely capillate, anterior commissure slightly uniplicate. Umbo short, sub-erect, truncated. Deltidial plates disjunct, pedicle-collar present, foramen submesothyrid, hinge-projections present, hinge-line terebratulid. Inner socket-ridges and crural-bases fused, cardinal process transverse; slight ridge dividing dorsal muscle-field; loop not seen.

Plesiotypes.

BB.13093–13101 from the Bracklesham Beds, Bed Fisher 21, Auversian; Selsey, Sussex.

[1] Previous Selsey material from near this level was referred to as Fisher 17 by correlation with the section at Whitecliff Bay, Isle of Wight (Fisher, 1862).

Description.

Early growth-stages are marked by about 16 to 20 obscurely nodose costae on each valve. A second series is seen to arise by intercalation at about 2 mm. shell-length, the nodose character soon being lost, and subsequent series are frequently inserted as growth proceeds. Internal shell-deposition converts these costae into capillae, of which the adult valve shows 70 or more marginally. The pedicle-opening in young shells is triangular, margined laterally by narrow interareas and anteriorly by hinge-projections; deltidial plates appear at a shell-length of about 2 mm., and are triangular and disjunct in the adult. Internally, the cardinalia are of the usual terebratulinid type, with fused crural bases and inner socket-ridges, and a small transverse cardinal process. The loop is not preserved but presumably was of the normal annular pattern for the genus. A faint ridge divides the brachial muscle-field; so far as can be made out the muscle-scars are similar to those of the London Clay *T. wardenensis* (Elliott, 1939).

Compared with *T. wardenensis*, the other British Eocene *Terebratulina* (Elliott, 1938), *T. edwardsi* is smaller in adult size, and has thinner test and more delicate cardinalia. It shows much finer, sharper, more closely-set and rather more numerous capillae at comparable sizes, originating mostly by intercalation rather than bifurcation. In outline it appears pentagonal with tapering posterior portion when compared with the bluntly-rounded London Clay species, and it is definitely uniplicate. In development the early nodose ornament is finer and disappears sooner except on the dorsal ears and corresponding ventral areas, foraminal structures appear earlier, and internal shell-deposition is thinner and relatively slower in converting costae to capillae. It is very similar to *T. putoni* (Baudon) as figured by Deshayes (1858; pl. 86, figs. 16–21), but attains a larger size (7–8 mm. shell-length compared with 5–6 mm.).

Argyrotheca spp.

Three specimens are referable to species of *Argyrotheca*; they are a pedicle-valve of *A. altavillensis* (de Morgan), a brachial valve of *A.* cf. *semicostata* (Baudon), and a worn fragment of the thick brachial hinge-area of a larger species than either of the other two. *A. altavillensis* was described from the Faluns à Cérites (Lutetian or Auversian) of the Cotentin (de Morgan, 1884; Cossmann and Pissarro, 1905). *A. semicostata* occurs in the Paris Basin Lutetian; the English specimen shows the thin test, hinge-structure and prominent shallow median sulcus of this species, as figured by Deshayes (1858; pl. 87, figs. 1–5), but the valve is damaged marginally and the subsidiary lateral sulci are not visible. The third *Argyrotheca* is indeterminable, but represents a larger species than the other two. All three are new to this country.

Notes on the British Eocene Fauna

British Eocene brachiopods were monographed by Davidson (1852, 1874) and listed up-to-date by Muir-Wood (1939). The fauna as now known is listed below: the stage names are after Wrigley and Davis (1937).

Upper Barton Beds.	Bartonian.	*Terebratula bartonensis* Muir-Wood.	Barton, Hants.
		Argyrotheca piperipyxis Elliott.	,,
Upper Bracklesham Beds.	Auversian.	*Terebratula cogidumni* Elliott.	Bracklesham, Sussex.
		Terebratulina edwardsi Elliott.	,,
		Argyrotheca altavillensis (de Morgan).	,,
		Argyrotheca cf. *semicostata* (Baudon).	,,
Lower Bracklesham Beds.	Lutetian.	*Discinisca davisi* Muir-Wood.	,,
	Cuisian.	*Lingula tenuis* J. Sowerby	,,
London Clay.	Ypresian.	*Lingula tenuis* J. Sowerby.	London & Hampshire Basins ; many localities.
		Discinisca insularis Muir-Wood.	Isle of Wight.
		Discinisca sp.	Lower Swanwick, Southampton, Hants.
		Terebratula hantonensis Muir-Wood.	Catisfield, Fareham, Hants, and near Southampton.
		Terebratulina wardenensis Elliott.	*Balanocrinus*-horizon of N.E. and S. Essex, London district and N. Kent ; Middle London Clay of London District ; Upper London Clay of Sheppey, Kent.
		Argyrotheca londinensis Elliott.	*Balanocrinus*-horizon of S. Essex.
Woolwich, Reading, and Blackheath Beds.	Sparnacian.	*Discinisca ferroviae* Muir-Wood.	South London, Croydon, N.W. Kent, W. Sussex.
Thanet Beds.	Thanetian.	Not known (doubtful records).	

Of the above list, all but four species are rare, being known from single or a few specimens each; only *Lingula tenuis*, *Discinisca ferroviae*, *Terebratula hantonensis*, and *Terebratulina wardenensis* are locally common. This rarity of English Eocene brachiopods was commented on by Walker (1889) who suggested that it might be due to unfavourable environmental conditions now represented by clastic sediments. In Japan and Australasia, however, rich brachiopod faunas occur in such beds. The present writer (Elliott, 1948), discussed the decline of brachiopods throughout geological time when compared with lamellibranchs and their progressive retreat to or survival in deeper water in many parts of the world, including the European area; he commented on the rarity, small size, and frequent stunting and immaturity of many of the French Eocene brachiopods as evidence of failure of these gregarious animals to establish themselves in local colonies in shallow-water environments with large molluscan populations. In this country the same phenomenon is traceable. *Terebratula hantonensis* at Catisfield and *Terebratulina wardenensis* at Sheppey, are exceptional in both indicating successful colonisation; the anchorage for the former was provided by local pebble-beds, and the latter species is seen, through successive horizons of the London Clay, to have migrated in what is now the London Basin area, always keeping to the stiff-clay, presumably deeper-water facies. *Lingula tenuis*, exceptional for a brachiopod in being a littoral form with quite different ecological preference, illustrates the same phenomenon in reverse, abounding in the presumed shallow-water sandy clays of the Portsmouth and Bognor area, with the individuals from elsewhere less common and often smaller. The rarity and small size of the *Argyrotheca* spp., *Terebratulina edwardsi*, etc., suggest unsuccessful attempts at colonisation in environments which at Barton and Bracklesham yield rich fossil faunas of mollusca and other invertebrates.

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Stone Stripes in Breckland, Norfolk

By A. S. Watt

(PLATE VIII)

Abstract

The usefulness of plants in directing the attention of the archaeologist to likely sites of human occupation is well known. Their value to the geomorphologist is indicated in the following example, which puts on record the existence of stone stripes in Breckland.

PARALLEL stripes of *Calluna*, mostly straight, occasionally curved, had been observed during the 'thirties in several places in Breckland. The distances between the stripes and their regularity suggested (and at the time were interpreted as) the ridging of earlier cultivation. All were on gentle slopes up to about 5°.

During the last war the soft sandy calcareous soil to a depth of about 18 to 24 inches was removed from part of one of these areas, exposing the unleached chalky boulder clay. By the time the photographs were taken (1952) erosion had brought into relief low ridges each about 5½ yards wide, consisting of hard compact boulder clay with numerous flints aggregated towards the middle line. The ridges were separated from each other by shallow furrows about 3½ yards wide of softer more sandy calcareous material with flints scattered on the surface. Moreover erosion has also brought into relief those flints in place down the middle part of the ridge. Some flints are oriented up and down the slope: many are not. But the point has not been statistically investigated.

The relation between the structure of the boulder clay and the surface vegetation is clear. Comparison with the intact soil shows that the soil is shallower over the ridge and the *Calluna* forms a stripe over the deeper soil mid-way between the ridges.

No section has been dug across the ridge and furrow, but a pit near Mundford, cut into a similar gentle slope, shows concentrations of flints with their long axes vertical. Frost phenomena of various kinds presumably referable to periglacial conditions, are abundant in chalk pits in Breckland. One of these is the differential sorting of material which on level ground is shown as polygonal areas, and on gentle slopes, as described here, as stripes. As far as observed there are no surface physical phenomena directly reflecting the differences in the boulder clay. Differential rates of leaching lead to differences in soil depth and concomitant chemical properties which are reflected in the pattern of the vegetation.

There is no movement of this kind now in the area which in the absence of the biotic factor (in the broadest sense) would support forest. And we have to ascend to altitudes above the tree limit in this country to find living stripes.

These stone stripes have now been seen by many visiting scientists, one of whom, Dr. D. Livingstone, of Dalhousie University, with experience of the arctic, describes the one on Thetford Heath as the finest fossil stone stripe he has seen.

THE BOTANY SCHOOL,
CAMBRIDGE.

EXPLANATION OF PLATE VIII

FIG. 1.—View across the stripes showing the ridges (white) and furrows (dark).

FIG. 2.—Middle of a ridge, showing the aggregation of flints.

CORRESPONDENCE

THE PLUTONIC HISTORY OF THE ASWAN AREA, EGYPT

SIR,—I would like to apologize for, and correct, two errors in the explanation of Text-fig. 1 of my paper " The Plutonic History of the Aswan Area, Egypt " (*Geol. Mag.*, 1954, vol. xci, pp. 484–497). The figure (p. 486) is a simplified and generalized sketch-map of the solid geology; outliers of Nubian Sandstone, late minor intrusions and small intrusions of New Granite have been omitted for the sake of clarity. The last part of the explanation of the figure should read:—

" Black: New Granite. Close hatching indicates increasing homogeneity of granite."

A. R. GINDY.

GEOLOGY DEPARTMENT,
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12th February, 1955.

REVIEW

DAS EISZEITALTER—Grundlinien einer Geologie des Quartärs; I, Die Allgemeinen Erscheinungen des Eiszeitalters. By P. WOLDSTEDT. vii + 374 pp., 136 figs., 4 tables. Ferdinand Enke, Stuttgart, 1954. Price 41 DM. (bound).

It is always useful to have a work by an expert who is an acknowledged world-wide authority on any subject. In this work Professor Woldstedt gives us a text-book which covers the basic principles of pratically all aspects of Pleistocene problems. An account of modern ice sheets and glaciers, their thickness, rate of movement, etc., is followed by details of glacial deposits. Here one may find a full account of the origin and character of till and the numerous outwash and periglacial deposits.

What is particularly valuable is a clear statement of his views on the glacial and interglacial successions in North Germany and N.W. Europe. He emphasizes the importance of the two marine transgressions of the Holstein Sea and the Eem Sea as marking the Interglacial episodes between the Elster-Saale ice advances on the one hand and Warthe Weichsel on the other, and places the gap between the Saale and Warthe as an interstadial.

The description of the drift succession in Britain occupies only half a page, but in this short space he puts his finger on many of our local problems. Some 65 pages are given to an account of the flora and fauna of the Pleistocene (including man and his implements).

The book ends with a summary of the various theories concerning the cause and duration of the ice age, treated in a very balanced way.

A useful and quite up-to-date bibliography of seventeen pages will be found to be of the greatest value.

W. B. R. K.

FIG. 1.

FIG. 2.

STONE STRIPES IN BRECKLAND.

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